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Abstract

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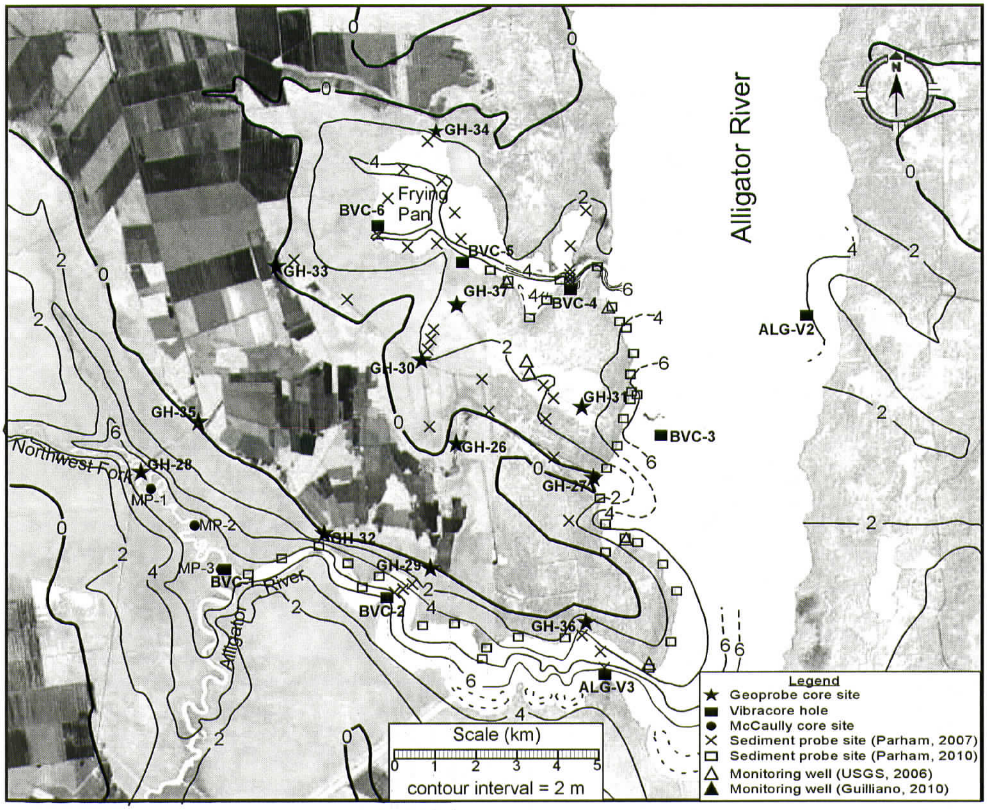

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ASTRONOMICAL TIDAL REGIME CHANGE AS A CONTROL ON THE HOLOCENE DEVELOPMENT OF AN ORGANIC-RICH COASTAL ZONE, NORTH CAROLINA, USA

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ABSTRACT

Sediment core and age data from the Alligator River tributary of Albemarle Sound, North Carolina suggest that coastal geomorphology-dependent variations in the astronomical tidal regime played a key role in the Holocene development of this presently nano-tidal estuarine system. The low lying, peatland-dominated study area is surrounded on three sides by drowned tributaries of the Alligator River estuarine system and underlain by a late Pleistocene topographic high. During the last glacial stage (MIS 2), drainage systems became established on the emergent terrain. As Holocene sea level rose, stream valleys became progressively inundated with concomitant up-gradient migration of wetlands. With open interchange between paleo-Albemarle Sound and the Atlantic Ocean, sea level was sufficiently high ca. 3 ka to produce bay/tidal ravinement inland of the present peatland shoreline. Gradual closure of the coastal barrier-island system (the Outer Banks) within the last approximately 2.5 ka resulted in decreasing astronomical tides, decreasing salinity, and marsh progradation seaward over the earlier Holocene ravinement surface with concurrent peat accumulation. Within the last ca. 1 ka, fresh to low-brackish marsh was replaced by forested wetlands in response to further barrier-island closure and elimination of astronomical tides as swamp-forest woody peat continued to vertically accumulate in response to rising base level.

INTRODUCTION

The response of a coastal region to rising sea level and changing hydrodynamics is complex with multiple interacting variables. Sediment supply is a major control on the evolution of many estuarine systems; surpluses tend to fill estuaries with sediment, whereas deficits typically cause estuarine shorelines to recede (Nichols, 1989; Brinson et al. 1995). Antecedent topography controls the character and aerial extent of transitional wetlands that form the leading edge of sea-level transgression. A gentle landslope promotes overland migration of wetlands in response to rising sea level while a steep landslope inhibits overland migration and results in wetland loss when there is low sediment supply (Brinson et al. 1995). Tidal conditions play an important role in determining the type and composition of estuarine wetlands (Fletcher et al. 1990). This study focuses on the effects of changing astronomical tidal regime on the Holocene evolution of an estuary and adjacent wetland system with a very low paleotopographic landslope and impounded behind an almost continuous sequence of barrier islands.

A record of the sedimentary and vegetation responses to late Quaternary changes in climate, sea level, and astronomical tidal dynamics is preserved at the Buckridge Coastal Reserve study area in the Albemarle Sound estuarine system (Fig. 1). Sea-level changes during the late Pleistocene were responsible for depositing and sculpting the underlying sediments upon which Holocene deposits, dominated by peat, developed and record the region's response to changes in climate and hydrodynamics.

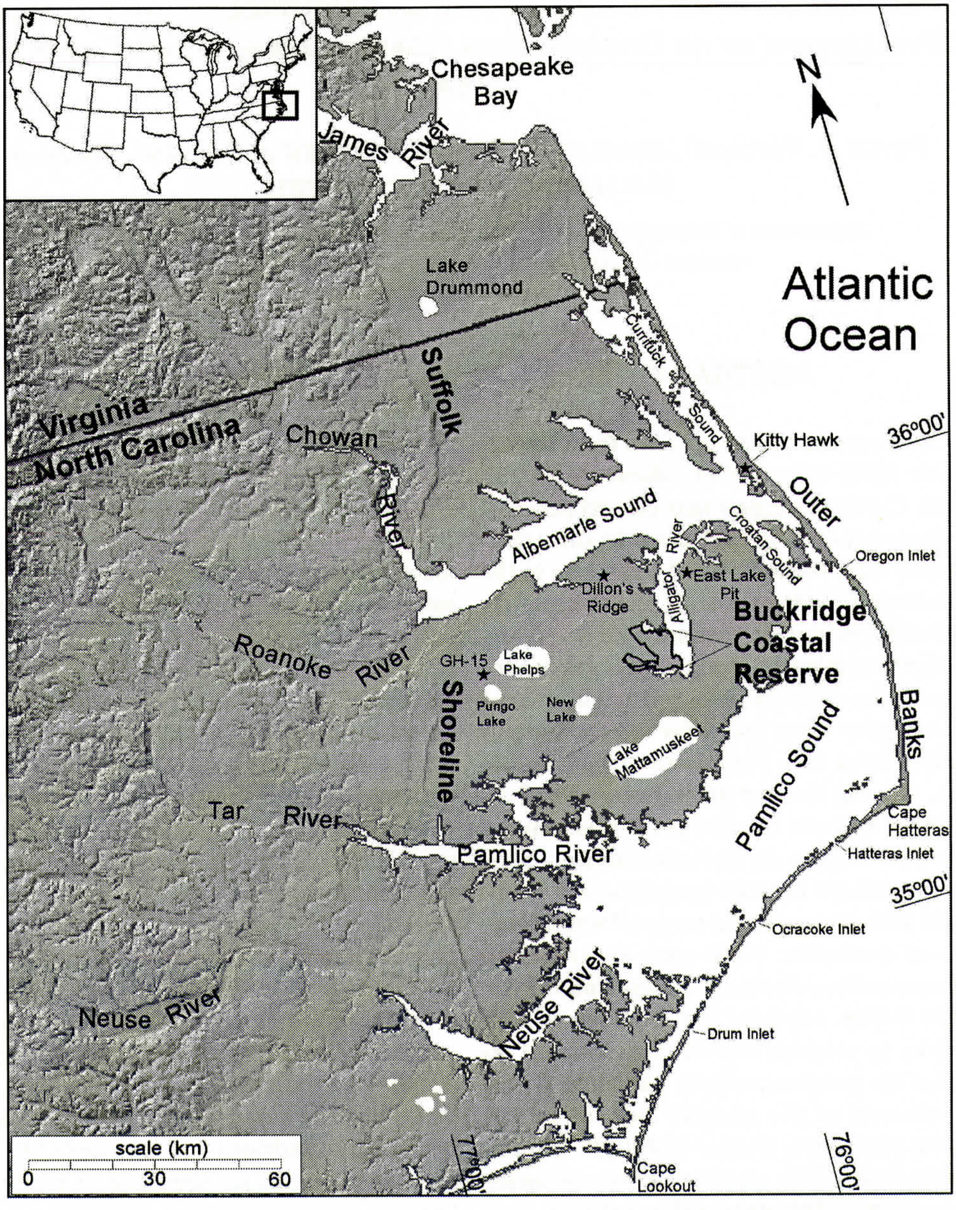

Figure 1. Shaded relief map (*http://nationalatlas.gov/*) of northeastern North Carolina and southeastern Virginia showing the location of the Buckridge Coastal Reserve study area. Major geographic features including the Suffolk Shoreline are also indicated. Core holes and pits from which additional stratigraphic and age data were derived are indicated by black stars. Insert map indicates location of Figure 1.

This study used lithologic and age data to investigate the late Quaternary sedimentary record of Buckridge Coastal Reserve (Fig. 1). The results demonstrate that the evolution of this coastal region during rising Holocene sea level was not characterized by progressive inundation, but rather was highly influenced by antecedent geomorphology, development of the Outer Banks barrier-island system, and changes in astronomical tidal regime.

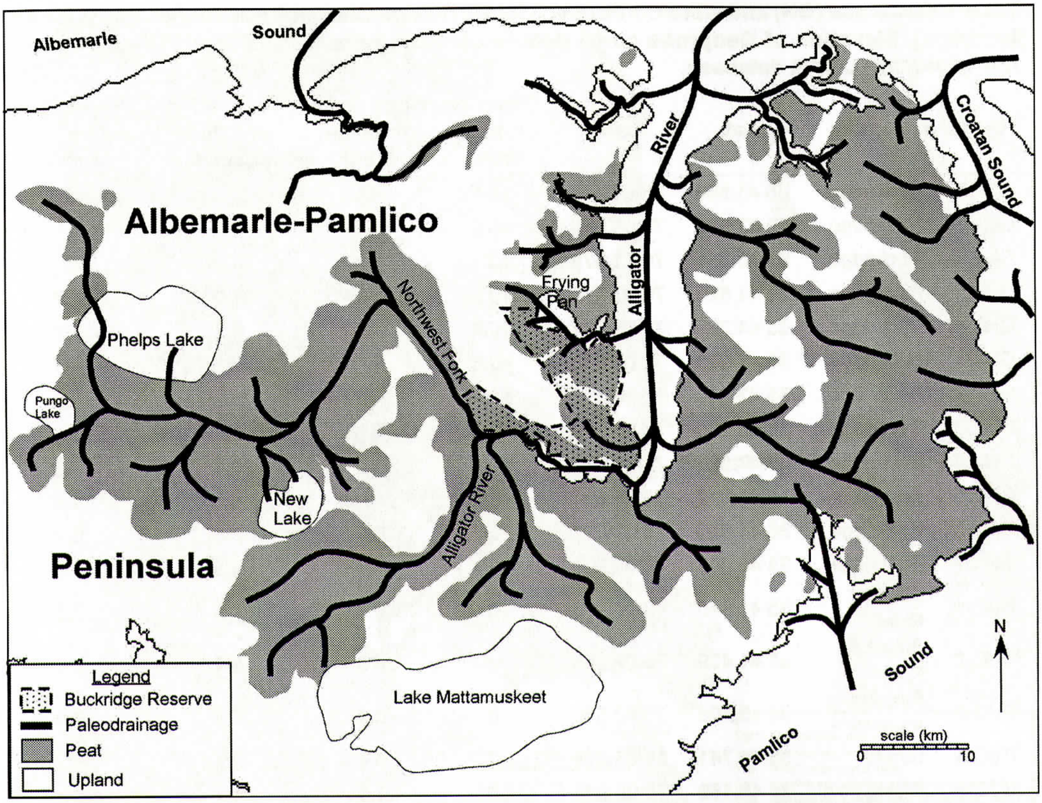

Figure 2. Map of the Albemarle-Pamlico Peninsula shows the distribution of peat (Ingram 1987) and paleo-drainage system of the Alligator River.

STUDY AREA

Albemarle Sound and Pamlico Sound to the south form the second largest estuarine system in the United States (Giese et al. 1985) and are separated from the Atlantic Ocean by the Outer Banks (Fig. 1). Clastic sediment input to the estuarine system is mainly through shoreline erosion of older deposits (Riggs and Ames 2003). The Albemarle-Pamlico estuarine system is connected with the Atlantic Ocean by four small inlets, all of which are south of the mouth of Albemarle Sound and Roanoke Island (Fig. 1). This results in negligible or nano-astronomical tides, domination by wind tides, and minimal salt water exchange producing fresh to low brackish conditions in the Albemarle Sound system (Giese et al. 1985; Riggs and Ames 2003). However, several studies suggest that for much of the Holocene, the system was very different from today. Culver et al. (2007) document initial flooding of southern Pamlico Sound beginning ca. 7000 calyr BP with salinity increasing to normal conditions by ca. 4000 to 3700 calyr BP. With the subsequent formation of barrier islands, Pamlico Sound became a semi-restricted estuary until ca. 1170 to 950 calyr BP when a portion of the barrier islands collapsed into submarine shoals (Culver et al. 2007). This caused southern Pamlico Sound to become dominated by marine conditions until ca. 550 to 420 calyr BP when the present barrier island system was re-established, returning brackish water, estuarine conditions to Pamlico Sound (Culver et al. 2007).

The Albemarle estuarine system and the Pamlico estuarine system were not hydrologically connected until ca. 200 yr BP and thus had different evolutionary histories (Riggs et al. 2000; Riggs and Ames 2003; Culver et al. 2007; Culver et al. 2008). Age and lithologic data indicate astronomical tidal conditions and tidalite

Table 1. Geoprobe (GH) and vibra-cores (BVC) for Buckridge Coastal Reserve (See Fig. 3 for core locations). Elevation of Geoprobe cores determined using the North Carolina Department of Transportation LiDAR database.

Core	Locality	Lat	Long	Elevation (m above MSL)	Water depth (m below MSL)	Penetration relative to surface (m)	Penetration relative to MSL (m)
GH-26	Buckridge	35 43.568	76 05.960	+0.8	NA	20.7	19.9
GH-27	Buckridge	35 43.043	76 03.653	+0.3	NA	8.2	7.9
GH-28	Buckridge	35 43.381	76 11.473	+0.3	NA	10.5	10.3
GH-29	Buckridge	35 41.828	76 06.570	+0.6	NA	8.5	7.9
GH-30	Buckridge	35 44.769	76 06.624	+0.4	NA	6.1	5.7
GH-31	Buckridge	35 43.993	76 03.690	+0.6	NA	7.3	6.7
GH-32	Buckridge	35 42.394	76 08.359	+0.3	NA	8.5	8.2
GH-33	Buckridge	35 46.188	76 08.908	+0.6	NA	6.1	5.5
GH-34	Buckridge	35 47.993	76 06.258	+0.3	NA	6.1	5.8
GH-35	Buckridge	35 44.012	76 10.521	+0.3	NA	4.9	4.6
GH-36	Buckridge	35 41.059	76 03.907	+0.6	NA	7.3	6.7
GH-37	Buckridge	35 45.565	76 05.990	+0.3	NA	4.9	4.6
BVC-1	Alligator River	35 41.945	76 10.163	0	0.3	7.3	7.3
BVC-2	Alligator River	35 41.439	76 07.339	0	0.6	6.4	6.4
BVC-3	Alligator River	35 43.608	76 02.556	0	0.4	8.9	8.9
BVC-4	Straits	35 45.741	76 03.938	0	0.4	9.2	9.2
BVC-5	Frying Pan	35 46.159	76 09.922	0	1.0	4.1	4.1
BVC-6	Frying Pan	35 46.756	76 07.407	0	1.2	4	4

accumulation in Albemarle Sound (Sager and Riggs 1998) and portions of Croatan Sound (Riggs et al. 2000) (Fig. 1) between 5500 and 2900 calyr BP with evidence of increasingly restricted tidal range following that time. Similar evidence along with paleontologic data from the Kitty Hawk area (Fig. 1) indicate open interchange between the paleo-Albemarle estuarine system and the Atlantic Ocean throughout most of the Holocene (Culver et al. 2008). Optically stimulated luminescence (OSL) ages for beach ridges in the Kitty Hawk area indicate that the inlet directly connecting Albemarle Sound with the Atlantic Ocean began to close ca. 2500 calyr BP and subsequently closed completely (Mallinson et al. 2008).

The study area is located at the southern end of the Gum Neck Peninsula between the Alligator River and Frying Pan Creek, flooded tributaries of the Roanoke/Albemarle drainage system (Figs. 1 and 2). With lowered sea level during the last glacial stage, the Alligator River system drained much of the Albemarle-Pamlico Peninsula region (Fig. 2). Paleo-Frying Pan Creek was a first-order tributary draining a relatively small area into the second-order Alligator River that flowed, in turn, into the third-order Roanoke River. Both the Alligator River and Frying Pan Creek drainage basins were progressively filled with peat in response to post-glacial climate warming and resulting rise in sea level. Today, the study area is dominated by forested wetlands that are underlain by peat ranging from 0 to over 7 m in thickness (Ingram 1987). Uplands are sparse and rarely exceed a meter in elevation. There is minimal sand or mud sediment input into the waters surrounding Buckridge Coastal Reserve.

Three important sedimentary processes are presently taking place in the study area in response to rising sea level and present hydrodynamic factors. 1) Within the freshwater forested

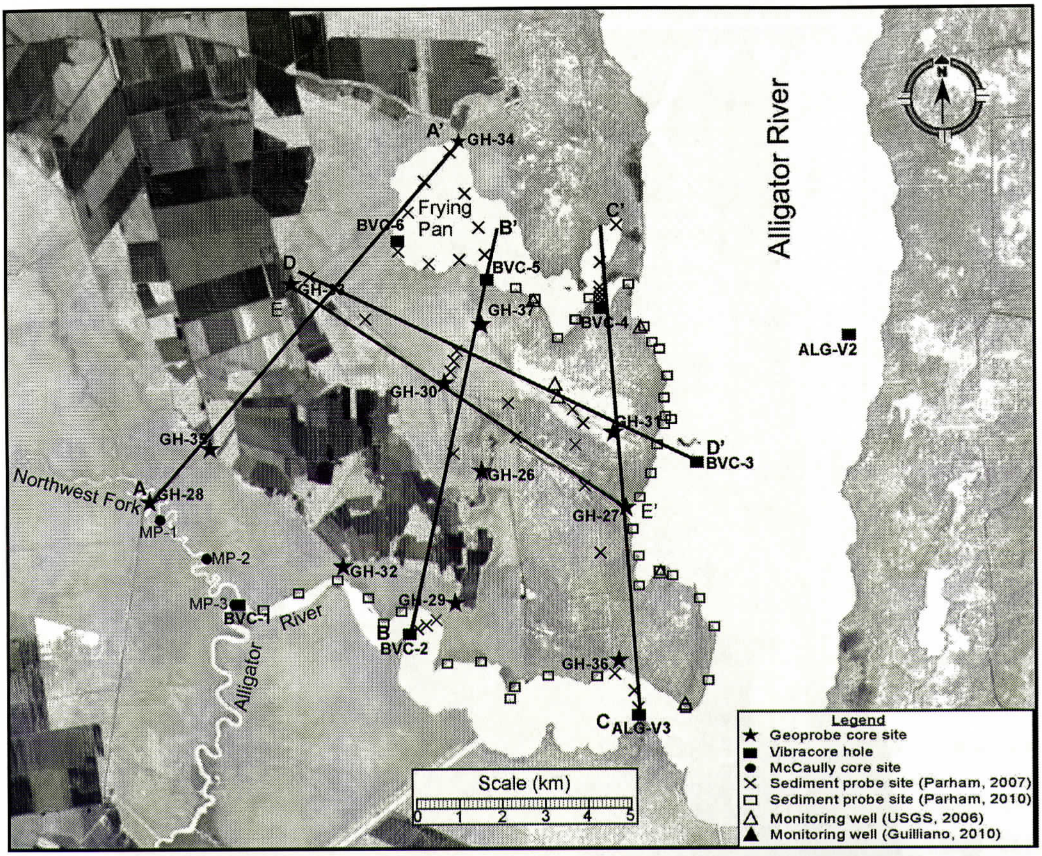

Figure 3. Map of the study area shows the locations of core holes, sediment probe sites, and transects (black lines) for which stratigraphic cross sections were prepared. B-B' cross section is shown in Figure 4. USGS (Ferrell et al. 2007) hydrologic survey wells also shown.

wetlands, peat is accumulating vertically. 2) Wetlands are migrating up-slope over uplands that are very low-gradient and low elevation. 3) Peat deposits along the estuarine shoreline are being eroded by bay ravinement.

METHODS

Eighteen cores, ranging in depth from 6 to 20 m, were acquired along five transects (Fig. 3). Twelve cores were obtained with a truck-mounted Geoprobe (direct push) coring device that used a hydraulically-driven hammer at a percussion rate of 30 Hz to obtain a 5 cm diameter continuous core in 1.2 m sections (Table 1 and Fig. 3). Eight 10 cm diameter continuous cores were obtained in the Alligator River and Frying Pan (Table 1 and Fig. 3) using a vibracore system. Peat thickness between cores was measured with a 9 m long, 1.3 cm diameter aluminum rod that was pushed down to the underlying mineral sediment contact. Samples to determine peat character and thickness were also obtained with a Macauly peat sampler. Latitude and longitude of each field site were determined using a Garmin 12-channel Global Positioning System (GPS). Surface elevations (+/- 25 cm) were determined using the North Carolina Light Detection and Ranging (LiDAR) database available from the North Carolina Department of Transportation (www.ncfloodmaps.com). Mean sea level (NAVD88) was used as a datum for all cores and stratigraphic sections.

Sediment cores were analyzed and logged both macro- and microscopically. Measure-

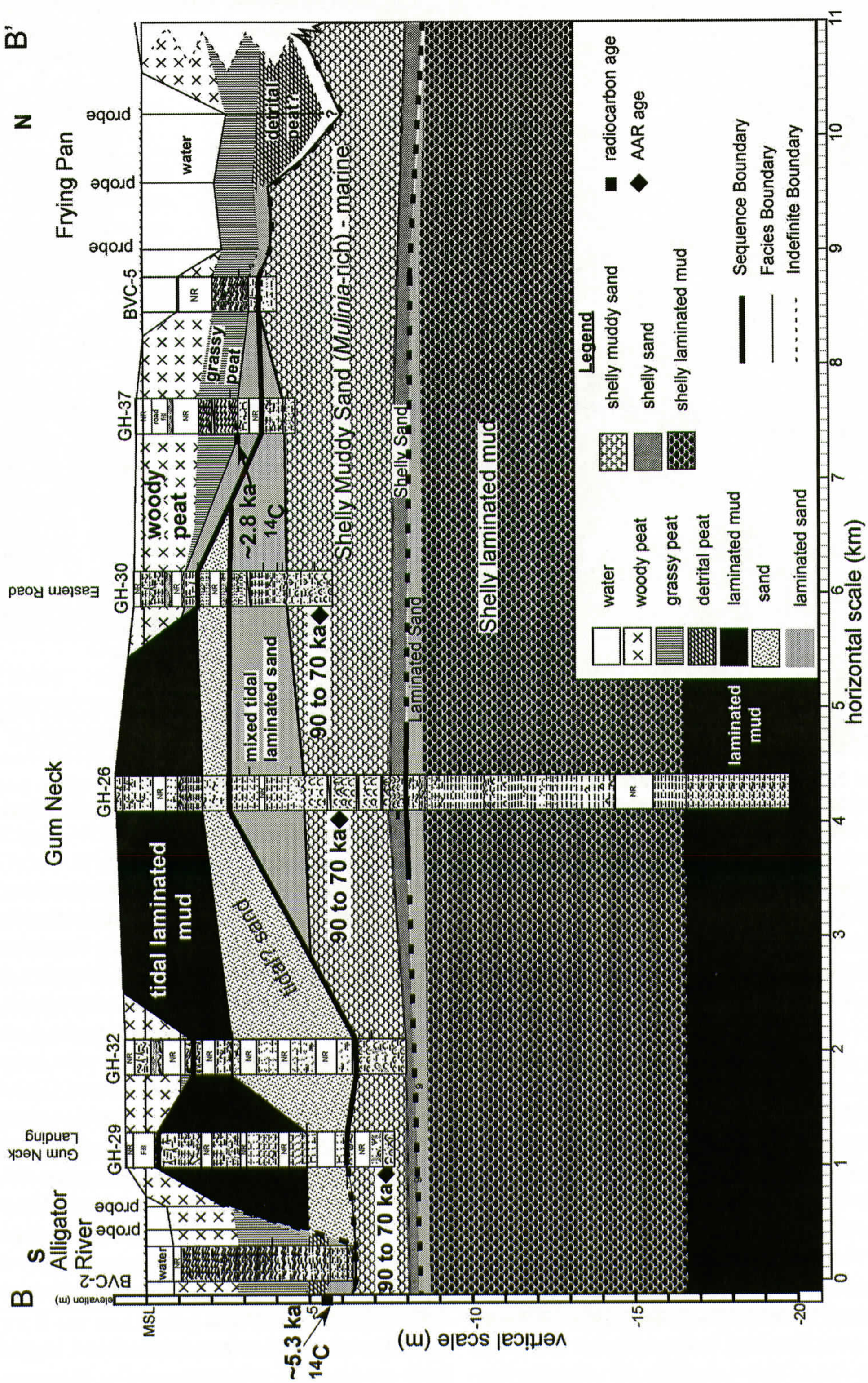

Figure 4. Stratigraphic cross-section of B-B' showing lithofacies and sequence boundaries. See Figure 3 for section location.

Table 2. Radiocarbon and amino acid racemization (AAR) age data for the Buckridge Coastal Reserve area (core locations in Fig. 3). Also shown are optically stimulated luminescence (OSL) age data from the Albemarle/Pamlico Peninsula.

Sample #	Locality	Latitude	Longitude	Depth bMSL (m)	Sample Type	Conventional Age (yr)	Age Error (+/- yr)	Calibrated 2σ age range (yr)
National Ocean Sciences Mass Spectrometry (NOSAMS) Radiocarbon Age Data								
GH-28-557	NW Fork Alligator River	35 43.381	76 11.473	5.57	muddy grassy peat	3290	20	3568 - 3462
GH-36-220	End Buck Island Rd.	35 41.059	76 03.907	2.20	grassy peat	1510	45	1447 - 1313
GH-37-300	N End Connector Rd.	35 45.565	76 05.990	3.00	grass in muddy sand	2790	35	2966 - 2788
BVC-2-561	Upper Alligator River	35 41.439	76 07.339	5.61	peaty mud	4590	40	5460 - 5263
BVC-3-725	Alligator River	35 43.608	76 02.556	7.25	detrital peat in mud	4710	35	5418 - 5322
BVC-4-672	Frying Pan Straits	35 45.741	76 03.938	6.72	detrital wood over cse sand	5050	40	5908 - 5710
BVC-4-708	Frying Pan Straits	35 45.741	76 03.938	7.08	detrital wood under cse sand	5200	40	6021 - 5897
BVC-4-768	Frying Pan Straits	35 45.741	76 03.938	7.68	detrital peat in mud	8910	65	10,214 - 9856
BVC-6-335	Western Frying Pan	35 46.756	76 07.407	3.35	laminated peaty mud	3450	35	3831 - 3635
Radiocarbon age data (S. Riggs, unpublished)								
ALG-V3-70	Big Bend Alligator River	35 40.671	76 03.062	0.70	muddy detrital peat	NA	NA	1660 - 1520
ALG-V3-97	Big Bend Alligator River	35 40.671	76 03.062	0.97	laminated detrital peat/ mud	NA	NA	1800 - 1640
ALG-V3-154	Big Bend Alligator River	35 40.671	76 03.062	1.54	detrital peat	NA	NA	2640 - 2460
ALG-V3-270	Big Bend Alligator River	35 40.671	76 03.062	2.70	muddy detrital peat	NA	NA	3670 - 3510
ALG-V3-370	Big Bend Alligator River	35 40.671	76 03.062	3.70	laminated mud with detrital peat	NA	NA	4470 - 4330
AAR age data						**Replicates**	**Estimated Age (ka)**	
GH-26-601	Buckridge interior	35 43.568	76 05.960	6.01	*Mulinia* (dwarf surf clam)	3	90 – 70	
GH-27-760	Grapevine Landing	35 43.043	76 03.653	7.60	*Mulinia* (dwarf surf clam)	4	90 – 70	
GH-29-760	Gum Neck Landing	35 41.828	76 06.570	7.60	*Mulinia* (dwarf surf clam)	3	90 – 70	
GH-30-533	Buckridge interior	35 44.769	76 06.624	5.33	*Mulinia* (dwarf surf clam)	6	90 – 70	
GH-33-400	Buckridge interior	35 46.188	76 08.908	4.00	*Mulinia* (dwarf surf clam)	2	90 – 70	
GH-33-510	Buckridge interior	35 46.188	76 08.908	5.10	*Mulinia* (dwarf surf clam)	6	90 – 70	
OSL age data from Albemarle/Pamlico Peninsula (Parham 2009)								
ELP-01	East Lake Pit, Dare Co.	35 53.122	75 57.434	8	shelly muddy sand	na	100.3+/-9.5	
ELP-05	East Lake Pit, Dare Co.	35 53.122	75 57.434	0.5	laminated mud and sand	na	66.9+/-6.7	
DR-2	Dillon's Ridge, Tyrrell Co.	35 55.776	76 11.252	0.2	laminated sand	na	62+/-12	
GH-18	New Lake, Hyde Co.	35 37.212	76 21.409	3.4	laminated sand	na	76.9+/-7.1	

Table 3. General character and interpreted depositional environments of lithofacies observed in the Buckridge Coastal Reserve study area. Patterns and shades correspond to lithofacies on stratigraphic cross sections.

General Character	Lithofacies	Depositional Environment
Peat	WP woody peat	swamp forest/shrub-scrub pocosin
	GP grassy peat	fresh to low-brackish marsh
	DP detrital peat	fluvial or estuarine
Non-Shelly Deposits	LM laminated mud	tidal fresh to low-brackish estuarine
	LS laminated sand	tidal fresh to low-brackish estuarine
	S sand	fluvial or tidal estuarine
Shelly Deposits	SSd shelly sand	shallow marine/shoreface
	SMS shelly muddy sand	shallow marine/open embayment
	SLM shelly laminated mud	low energy, brackish to marine, outer estuarine or open embayment

Observed Depositional Succession

ments were adjusted to allow for core compaction. Physical properties of sediments were characterized with respect to color, texture, grain size, and mineralogy using Folk's (1974) protocol. Fossil plant material was described and divided into three categories based on the characteristics of dominant constituents: *in situ* woody peat, *in situ* grassy peat, and detrital peat. Five lithostratigraphic cross sections were prepared using Canvas software. Because of their similarity, only cross-section B-B' (Figs. 4 and 5) is included here as illustrative of the other four.

Fifteen peat subsamples were analyzed for plant macrofossils. Samples were prepared and then identified according to Watts and Winter (1966) and Peteet (1986). Macrofossils were compared to the seed collection at Lamont-Doherty Earth Observatory and the guides of Martin and Barkley (1961), Fernald (1970), Montgomery (1977) and L6vesque et al. (1988). The volume of visible charcoal fragments was estimated.

Nine peat samples were selected for radiocarbon age dating and analyzed by the National Ocean Sciences Accelerator Mass Spectrometer (NOSAMS) Facility in Woods Hole, MA. The conventional radiocarbon age estimates reported by NOSAMS were calibrated using Calib 6.0.1 software (Stuiver et al. 2010) and the IntCal09 terrestrial calibration curve (Reimer et al. 2009). Additional radiocarbon age data (S. Riggs, unpub.) from the Alligator River and processed by Beta Analytic were also used for this study.

Mercenaria and *Mulinia* shells were collected from selected cores at elevations determined to be representative of particular lithologic

Table 4. Plant macrofossil analysis results.

Sample #	Lithofacies	Latitude	Longitude	Depth bMSL (m)	Sample Type	Depositional Environment
GH-28-570	Detrital peat	35 43.381	76 11.473	5.7	Roots in mud, *Nyssa* seed, *Carex trigonous* seed	probably fresh water, swamp forest
GH-30-120	Woody peat	35 44.769	76 06.624	1.2	Lots of grass epidermis, charcoal, *Polygonum* seed, roots, wood, organic-rich mud	brackish or fresh water
GH-30-145	Woody peat	35 44.769	76 06.624	1.45	Epidermis, charcoal	brackish or fresh water
GH-31-235	Grassy peat	35 43.993	76 03.690	2.35	Detrital grass, wood, charcoal	brackish or fresh water
GH-36-200	Grassy peat	35 41.059	76 03.907	2.0	Shredded epidermis and roots, *Cladium* seed, *Scirpus* seeds	possibly brackish water
GH-36-235	Grassy peat	35 41.059	76 03.907	2.35	Epidermis of grass, roots, fungal sclerota, charcoal	brackish or fresh water
GH-37-250	Grassy peat	35 45.565	76 05.990	2.5	Fine rootlets, hyphae, *Betula* twig, charcoal, insect parts	probably fresh water
GH-37-290	Grassy peat	35 45.565	76 05.990	2.9	Sedge stems, charred sedge stems, charcoal	brackish or fresh water
GH-37-310	Grassy peat	35 45.565	76 05.990	3.1	Bark, sedge stems, insect bristle, no charcoal	brackish or fresh water
BVC-3-350	Grassy peat	35 43.608	76 02.556	3.5	Much sedge with many sedge nodes, *Cladium* seeds, *Compositae* seed, *Scirpus* seed, charcoal, insect (beetle?) wings	possibly brackish water
BVC-3-635	Detrital peat	35 43.608	76 02.556	6.35	Wood, grass stems	probably fresh water
BVC-3-725	Detrital peat	35 43.608	76 02.556	7.25	3-needle and 2-needle pine needles, bark	fresh water
BVC-5-325	Grassy peat to woody peat	35 46.159	76 09.922	3.25	Wood fragments, charcoal, quartz sand, grass epidermis?	probably fresh water
BVC-6-340	Woody peat	35 46.756	76 07.407	3.4	Wood, twigs (*Betula*), charcoal, *Scirpus* seed	probably fresh water
BVC-6-355	Laminated mud/wetland soil	35 46.756	76 07.407	3.55	Wood, rootlets	?

units. Sample selection was determined primarily by availability of suitable, minimally-altered specimens in quantities sufficient for dating. Forty-four shell samples were submitted to the Amino Acid Geochronology Laboratory, Northern Arizona University for AAR age estimation. The resulting age data were interpreted by John Wehmiller, University of Delaware.

CHARACTERIZATION OF LITHOFACIES AND INTERPRETED PALEOENVIRONMENTS

Lithofacies observed in the study area are listed in Table 3.

Non-shelly deposits

Peat is confined to the Holocene section and thickness varies from < 10 cm along the fringes of paleo-valleys to ca. 6 m in the deepest sections. Peat is categorized based on its dominant constituents and includes woody peat, grassy peat, and detrital peat.

Woody peat is dominated by root fibers and wood. Inorganic mud and sand are rarely present. Root fibers are generally elongate, non-abraded, and vertically oriented indicating that peat accumulated *in situ* rather than having been transported (detrital) and then deposited. Roots and logs are common and may exceed 10 cm in diameter. Because of the dominance of woody material and evidence of *in situ* accumulation, woody peat is interpreted to represent peat formation in a forested wetland or shrub-scrub pocosin-like environment similar to that which covers the study area today. Woody peat generally forms the upper 2 m of sediment throughout the peat lands of the study area (Figs. 3 and 4).

Grassy peat consists of soft, fibrous, bladed grassy material. Inorganic mud is a common accessory constituent and typically increases down section from trace amounts to as much as 40%. Most plant stems and root fibers are vertically oriented indicating *in situ* accumulation. Broad-bladed *Typha* (cattail) fragments with obvious perpendicular partitions within the leaves are occasionally present. Grassy peat typically occurs between 2 and 5 m below mean sea level (bMSL) (Fig. 4). Plant macrofossil analysis of grassy peat (Table 4) along with a lack of observed foraminifera and the presence of *Typha* suggest formation in a fresh to low-brackish marsh environment (McNaughton, 1966; M. Brinson, pers. comm.).

Detrital peat contains up to 40% mud and trace amounts to 5% very fine to fine grained sand. Detrital peat is typically interlayered/interlaminated with organic-rich mud and/or sand. Plant material is horizontally oriented, compressed, and dominated by rounded wood fragments, leaves, and seeds. The rounded character and horizontal orientation of the constituents of detrital peat indicate that they were transported and abraded by current or wave action prior to deposition. Similar deposits are found along the modern eroding peat shorelines within the study area. Detrital peat occurs below grassy peat in paleo-valleys associated with both the Alligator River and Frying Pan drainage systems at depths up to 7.8 m bMSL (Figs. 3 and 4). Interpretation of the depositional environment is largely based on stratigraphic context. Detrital peat encountered within the basal portion of thick peat sections potentially represents fluvial deposition in paleo-stream channels. However, the widespread occurrence of detrital peat elsewhere in the study area suggests shoreline erosion of pre-existing peats and re-deposition in an estuarine system.

Laminated mud consists of laminated to layered, very fine to medium sand, sandy mud, and mud. Mica is an accessory constituent ranging from 1% to 5%. Up to 20% mica is present in some laminae/layers. Trace amounts to 2% of fine plant detritus, wood fragments, and heavy minerals are typically present. In cores, laminated mud manifests coarsely interlayered bedding with thicker mud laminae and thinner sand laminae. Outcrops of late Pleistocene laminated mud in the northeastern North Carolina region display lenticular to wavy bedding structures indicating deposition in a muddy intertidal environment (Reineck and Singh 1973; Mallinson et al., 2008).

Laminated sand is dominated by medium grained sand with mud laminae. It generally

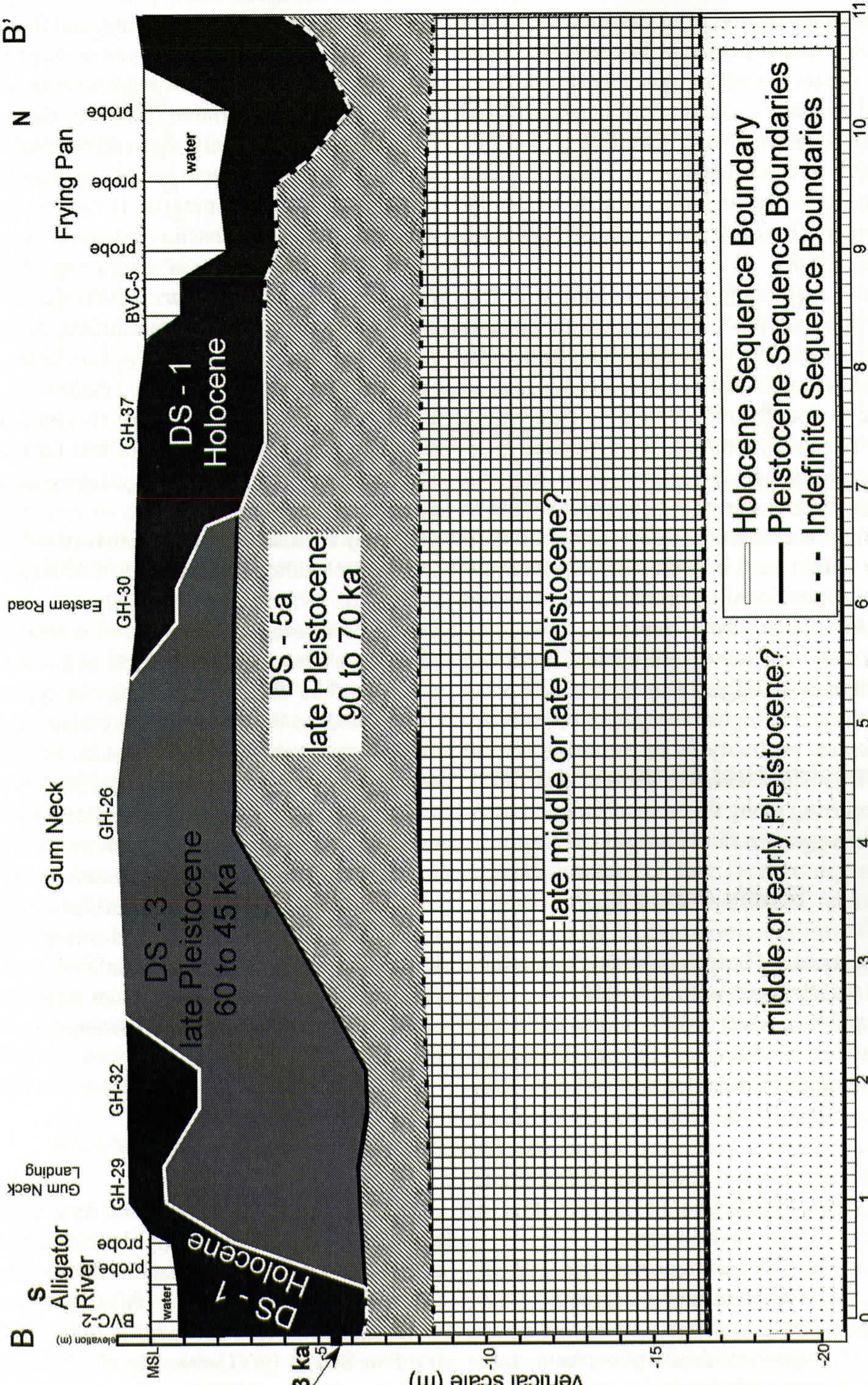

Figure 5. Depositional sequence (DS) relationships along transect B-B'. See Figure 3 for section location.

contains <10% coarse sand and <2% mica. Trace amounts of fine plant detritus are associated with muddier horizons. In cores, laminated sand is characterized by coarsely interlayered bedding structures. Outcrops of laminated mud in the northeastern North Carolina region display wavy to flaser bedding structures indicating deposition in an inter-tidal to sub-tidal environment (Reineck and Singh 1973; Mallinson et al., 2008).

Sand is medium grain quartz sand with less than 20% coarse sand, less than 4% very coarse sand, less than 2% mica and heavy minerals, and trace amounts of fine plant detritus. Up to 8% fine to coarse-grained mica occurs within occasional muddy laminae. Well-rounded granule size gravel is also occasionally present. Sand may be massive or may alternate between coarser and finer layers. Heavy minerals form laminae within sand in some sections. Interpretation of depositional environment for the sand lithofacies is largely based on stratigraphic context. Laterally extensive beds of sand occurring upon interfluves suggest that these sand bodies are the product of higher energy tidal or sub-tidal processes, probably in an estuarine setting (Reineck 1972; Finkelstein and Ferland 1987). Sand occurring along the base of incised paleochannels suggests a fluvial origin.

Shelly deposits

Shelly sand is comprised of fine to medium grained quartz sand with 2 to 10% shell and shell fragments, trace to 10% coarse sand, and trace amounts to 1% heavy minerals. Shell fragments are typically rounded and include *Mulinia*, *Ensis*, *Mellita*, *Dosinia*, Tellinidae, Cardiadae, *Dentalium*, and ostracods. The rounded and reworked character of shell material suggests deposition in either a marine shoreface or shoal environment.

Shelly muddy sand consists of shelly, slightly muddy, fine to medium grained quartz sand with <20% coarse sand, <5% very coarse sand, <3% heavy minerals/phosphate, trace amounts of mica and glauconite, and 2 to >25% fossil material dominated by mollusk shells. Articulated bivalves and other evidence of *in situ* burial are occasionally present, but typically disarticulation, slight rounding, and fragmentation of shells indicate some degree of reworking. Shells range from well-preserved to white, chalky, and friable. Black or dark grey coloration of some shells suggests original burial in mud and subsequent reworking into shelly muddy sand. Fossil material is dominated by *Mulinia*. Accessory fossil constituents may include Tellinidae, *Ensis*, *Parvilucina*, *Melita*, *Nucula*, *Terebra*, *Anadara*, *Olivella*, *Oliva*, *Nassarius*, *Urosalpinx*, foraminifera, bryozoa, reworked *Crassostrea*, *Spisula*, Cardiadae, *Divaricella*, barnacles, *Chione*, *Polinices*, and *Crepidula*. Foraminiferal analysis (Parham et al. 2007; Culver et al. 2008) of this lithofacies indicates accumulation in a shallow marine open embayment.

Shelly laminated mud consists of interlaminated, fine sandy mud, muddy fine sand, and mud with <3% mollusk shell/shell fragments and trace amounts of medium to coarse sand, phosphate/heavy minerals, mica, and fine plant detritus. The fossil assemblage is typically dominated by *Mulinia* but may also include Tellinidae, *Ensis*, bryozoa, *Nucula*, *Nuculana*, ostracods, *Abra*, and foraminifera. Shell preservation ranges from pristine articulated bivalves to chalky shell material to indeterminate shell fragments. Shells are concentrated in layers/laminae, consisting of small or immature specimens. Shelly laminated mud is interpreted to have been deposited in either an estuarine setting with regular tidal interchange with marine waters or a shallow marine environment.

AGE RESULTS

Radiocarbon

Radiocarbon age analysis of nine organic samples indicated four fairly distinct intervals of peat and peat-rich sediment accumulation (Table 2). 1) Basal (7.68 m bMSL) peaty mud from a paleo-stream meander underlying Frying Pan Straits (BVC-4 in Fig. 4) yielded the oldest age range of 10,214 to 9856 cal yr BP. 2) Detrital organics and peaty mud between the depths of 5.6 and 7.2 m from cores in Frying

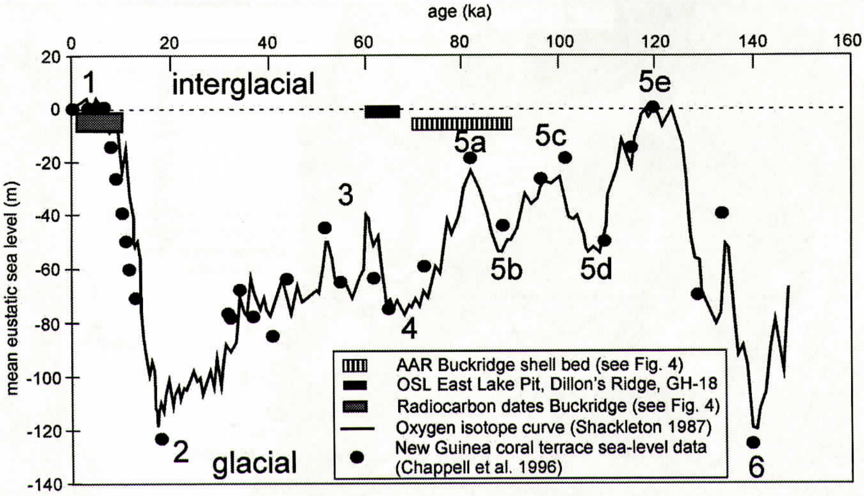

Figure 6. The timing and vertical occurrences of age-dated samples from Buckridge are compared with an oxygen isotope curve (Shackleton 1987) that has been calibrated with sea-level data from New Guinea to represent changes in eustatic sea level over the last ca. 150 ka (Chappell et al. 1996). Numbers along curve correspond with marine isotope stages.

Pan Straits (BVC-4) and Alligator River (BVC-2 and -3) (Fig. 3) produced similar ages between 6021 and 5263 cal yr BP. Note that, because detrital material has been reworked and transported, radiocarbon ages may not reflect the time of deposition (i.e. depositional age may be younger). 3) Muddy peat (transitional between peat above and dominantly inorganic sediment below) from cores GH-28 (NW Fork Alligator River) and BVC-6 (western Frying Pan) (Table 2 and Fig. 3) ranged in age from 3831 to 3462 cal yr BP. 4) In two cores from the Buckridge interior (GH-36 and -37 in Fig. 3), grassy peat between 2.2 and 3.0 m bMSL produced age estimates between 2966 and 1313 cal yr BP (Table 2).

Amino acid racemization

Seven *Mulinia* samples from five cores (GH-26, GH-27, GH-29, GH-30, and GH-33 in Table 2) were submitted to the Amino Acid Geochronology Laboratory, Northern Arizona University for AAR age analysis. All samples are from shelly muddy sand (Fig. 4) and are of the same approximate age. AAR analysis results indicate that the shelly muddy sand facies was deposited between 90 and 70 ka (Table 2) which correlates temporally with marine deposits at similar depths throughout the region (Riggs et al. 1992; Mallinson et al. 2008; Wehmiller et al. 2010; Table 1).

Optically stimulated luminescence

Optically stimulated luminescence (OSL) age estimates for late Quaternary sea-level highstand deposits in the region, including age data from the eastern portion of the Albemarle-Pamlico Peninsula (Table 2), dominantly correlate with marine isotope stages (MIS) 5a and 3 (Mallinson et al. 2008; Parham 2009). Age assignments for Buckridge deposits, where direct OSL ages were not obtained, are based on these regional ages, depths, and correlative lithofacies.

DEPOSITIONAL SEQUENCES

Depositional sequences (DS) are each comprised of multiple lithofacies (Fig 4). Based on core and age data, portions of five depositional sequences occur within the upper 20 m of deposits underlying Buckridge Coastal Reserve (Fig. 5). Because of the relatively shallow nature of core holes in the Buckridge area, an en-

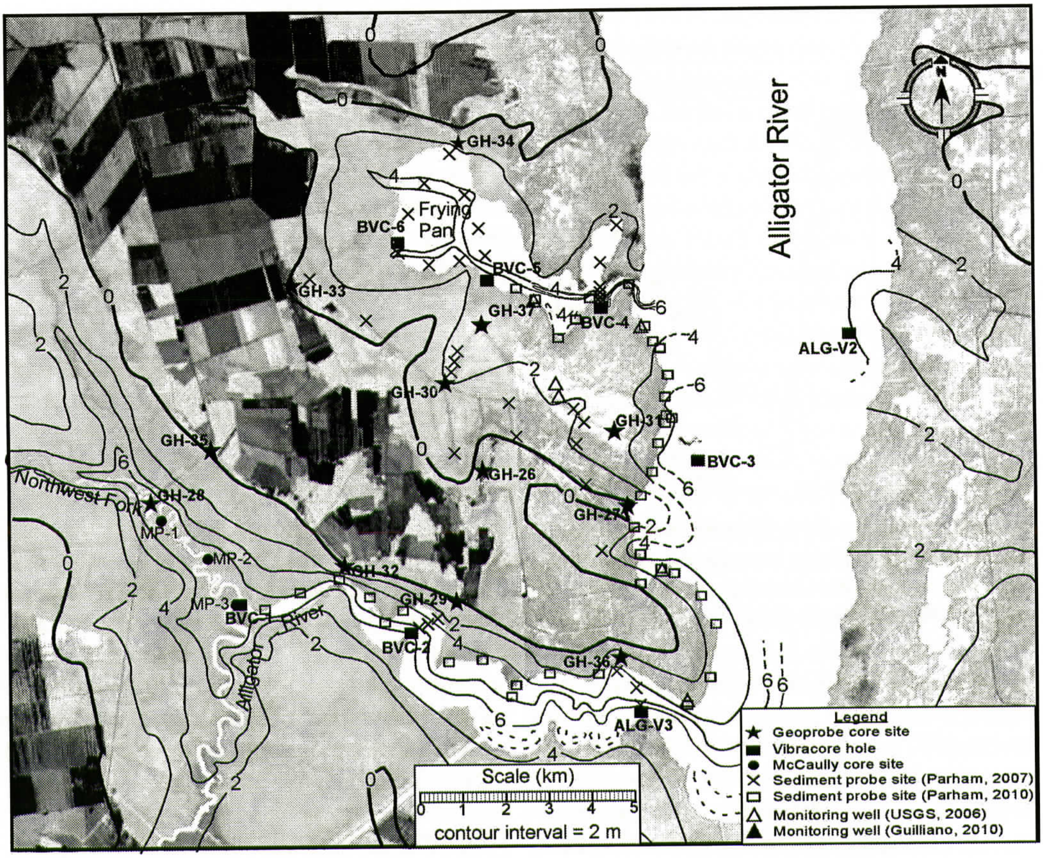

Figure 7. Structure contour map of the southern Alligator River area shows the paleo-topography of the pre-peat surface. Data are based on this study and Ingram (1987).

tire depositional sequence was not recovered in any single core.

Pleistocene framework

Portions of what are interpreted to represent two depositional sequences occur below –8.5 m in core-hole GH-26 (Figs. 4 and 5). Based on regional OSL and AAR age data (Mallinson et al. 2008; Parham 2009; Wehmiller et al. 2010), the lowermost sequence is likely of early or middle Pleistocene age and the overlying one probably late middle to late Pleistocene. Both of these deeper depositional sequences are dominated by laminated mud and shelly laminated mud (Fig. 4).

DS-5a (Fig. 5) correlates well with other regional stratigraphic data (Wehmiller et al. 2010) and AAR age estimates indicate an age range of 90 to 70 ka (Table 2, Fig. 4). The lower depositional sequence boundary for DS-5a in core GH-26 (Fig. 4) is characterized by an abrupt shift in lithofacies from organic-rich laminated sand (below) to shelly sand (above) and is interpreted to represent an open-ocean shoreface ravinement surface. DS-5a consists of an upward succession from shelly sand to shelly muddy sand to laminated sand (Fig. 4).

DS-3 extends from the upper surface of DS-5a up to the modern land surface in upland areas and to the base of the Holocene deposits beneath wetland and estuarine areas (Fig. 5). The basal depositional sequence boundary of DS-3 is characterized by an abrupt shift from shelly muddy sand or laminated sand below to sand with mud rip-up clasts above. DS-3 consists of an upward succession from burrowed sand to laminated mud (Fig. 4). No age data were ac-

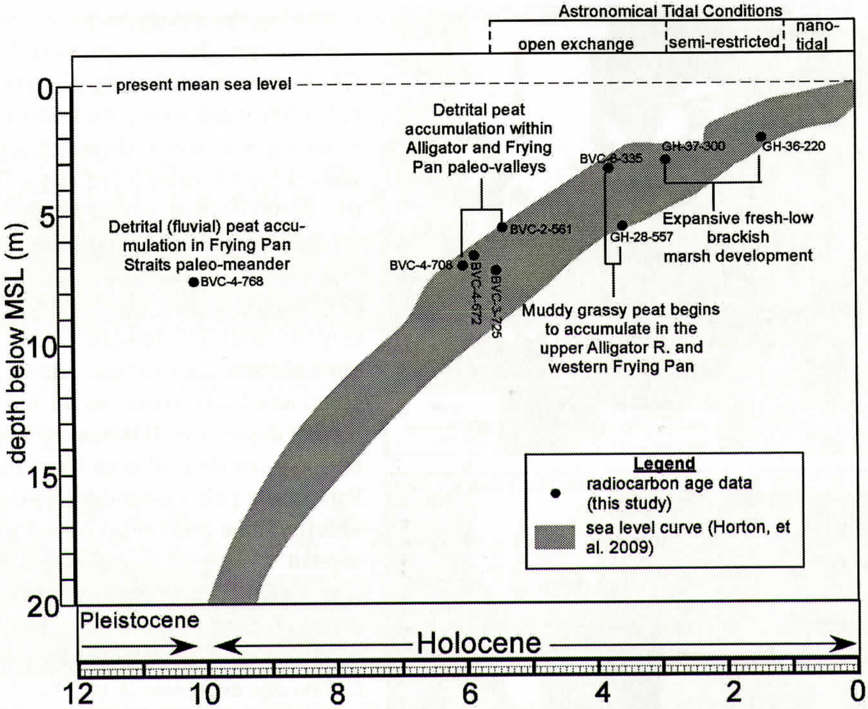

Figure 8. Plot of radiocarbon ages and sample depths from the Buckridge study area with a relative sea-level curve (grey field from Horton et al. 2009) based on sea-level index points and terrestrial and marine limiting data from northeastern North Carolina. Astronomical tidal conditions are based on data from Sager and Riggs (1998), Riggs et al. (2000), and Culver et al. (2008).

quired for DS-3 within the Buckridge study area. However, OSL analysis (Parham 2009) of similar deposits from the Albemarle/Pamlico Peninsula (East Lake Pit and Dillon's Ridge in Fig. 1) at depths ranging from -0.6 to +1.5 m (relative to MSL) produced ages from 67 to 62 ka (Table 2) suggesting a similar age for these deposits at Buckridge.

Holocene

DS-1 (Fig. 5) occurs in the paleo-valleys of the Alligator River and Frying Pan tributary creek. In the deeper portions of these paleo-valleys, below ca. 4 m bMSL, DS-1 grades upward from basal laminated sand and sand into detrital peat that, in turn, is overlain by grassy peat and woody peat (Fig. 4). Where the paleo-valley floor is less than ca. 4 m bMSL, DS-1 is comprised of an upward succession from laminated sand or sand to grassy peat to woody peat (Fig. 4). Woody peat is the dominant peat type throughout the interior of the reserve (Fig. 4). Radiocarbon age estimates for DS-1 range from ca. 10 ka to 1.3 ka. (Table 2).

DISCUSSION

Pleistocene deposits at Buckridge form the framework upon which the Holocene organic-rich coastal system evolved. Uppermost Pleistocene deposits consist of both shelly and non-shelly lithofacies that comprise DS-5a and DS-3 (Figs. 4 and 5). AAR age estimates indicate that DS-5a correlates with the MIS 5a sea-level highstand ca. 80 ka (Fig. 6). DS-3 is stratigraphically constrained to a younger sea-level highstand and is correlated with MIS 3 (Fig. 6) based on regional stratigraphic and age data (Mallinson et al., 2008; Parham, 2009; Wehmiller et al. 2010).

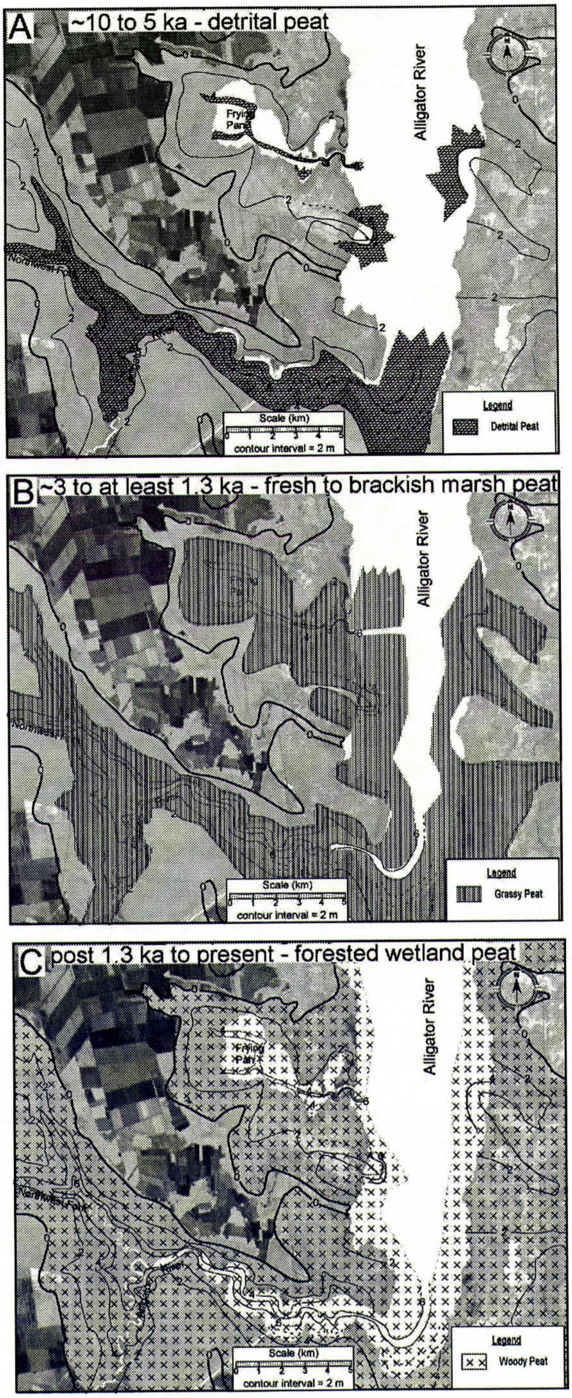

Figure 9. Evolutionary development of peaty deposits in the southern Alligator River area based on age data and peat stratigraphy.

During the emergent period associated with the last glacial stage, both DS-5a and DS-3 were modified by fluvial erosion, subaerial weathering, and soil formation resulting in a stream-dissected topography with at least 8 m of relief (Fig. 7). This is the surface upon which Holocene sediments (dominantly peat) accumulated. Radiocarbon age estimates are fairly broadly distributed within the Holocene section (Fig. 8; Table 2). However, early and middle Holocene ages for the detrital peats are considered unreliable as an indicator of time of deposition. If the ca. 10 ka radiocarbon age for detrital peat from the Frying Pan Straits paleo-meander (BVC-4 in Fig. 3) reflects the true timing of deposition, the deposit is likely of fluvial origin because it was well above relative sea level at that time (Horton et al. 2009) (Fig. 8). The depths of the four middle Holocene radiocarbon age estimates (6 to 5 ka) for detrital peat correspond with relative sea level at that time (Fig. 8). However, it is possible that this detrital peat was redeposited as the product of bay ravinement later than the radiocarbon ages suggest. Bottom sediments in the modern Alligator River system are largely comprised of plant detritus that eroded from peat shorelines. Because radiocarbon ages are based on the time of an organism's death, dates for detrital peat tend to be older than the actual time of deposition with an error up to several thousand years. The approximate distribution and depth of detrital peat encountered in the study area are shown on Figure 9A.

Ages for *in situ* peat are considered reliable indicators of depositional time and are reflective of the evolutionary history during the late Holocene. From ca. 3.5 ka to ca. 1.3 ka, grassy (fresh to low-brackish marsh) peat accumulation buried the basal detrital peat as the marsh extended onto low interstream areas between 2 to 4 m below present sea level (Fig. 9B). Grassy peat is overlain by woody peat reflecting a transition from open marsh to forested wetlands during the last approximately 1000 years (Fig. 9C).

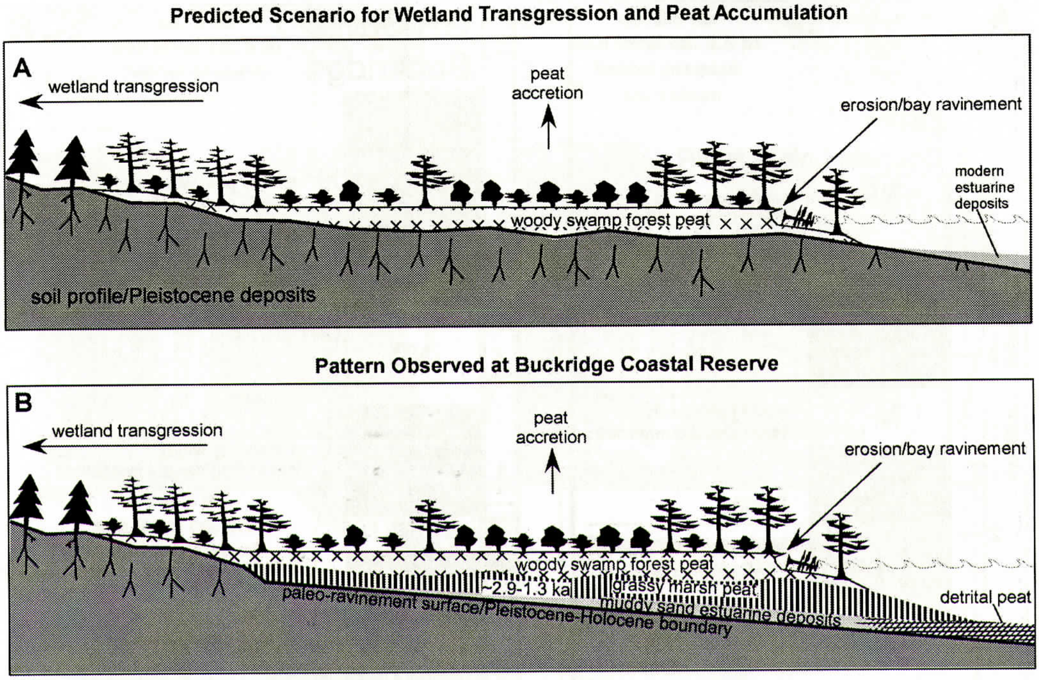

Figure 10. Schematic models comparing the stratigraphic patterns expected (Panel A) with those observed at Buckridge Coastal Reserve (Panel B).

The peat succession and character of underlying deposits at Buckridge are very different than expected. The hypothesis for the Holocene evolution of the study area expected a gradual conversion of low-slope uplands to wetlands in response to rising sea level with attendant peatland expansion through paludification and vertical peat accumulation largely in-pace with sea-level rise rates (Fig. 10A). In vertical section, a rooted soil profile on top of the Pleistocene deposits was expected to be overlain by woody forested wetland peat (Fig. 11A). What was actually observed in most inland cores was no evidence of soil beneath the Pleistocene-Holocene boundary and an overlying succession from 1) increasingly grassy, muddy sand to 2) grassy marsh peat to 3) woody forested wetland peat (Figs. 10B and 11B). This raises the question: how did this succession develop under the relatively high and slowly rising sea-level conditions of the last 3 to 3.5 ka?

Micropaleontologic and sedimentologic data (Culver et al. 2008) indicate that there was a greater astronomical tidal signal within the Albemarle Sound estuarine system earlier in the Holocene with relatively open exchange with Atlantic Ocean water between ca. 5.5 ka and 2.9 ka (Figs. 8 and 12A) (Sager and Riggs 1998) and progressive closure of the barrier island system between then and present (Fig. 12B and C). OSL data (Mallinson et al. 2008) indicate that relative sea level was high enough between 3.0 and 2.5 ka (Fig. 8) to produce beach ridges west of the present shoreline at Kitty Hawk (Fig. 12B). Lithologic and radiocarbon age data from Croatan Sound suggest open to semi-restricted estuarine conditions between 5.2 and 3.2 ka, between 3.0 and 1.1 ka, and then relatively closed estuarine conditions from ca. 1.1 ka to present (Figs. 8 and 12) (Riggs et al. 2000).

Data from the salt-marsh deposits adjacent to Croatan Sound (Horton et al. 2009; Kemp 2009; Kemp et al. 2009) indicate sea level was ca. 3 m below present approximately 3 ka (Fig. 8) when the Kitty Hawk beach ridges were being formed (Mallinson et al. 2008). With sea level within 3 m of present and relatively open estuarine con-

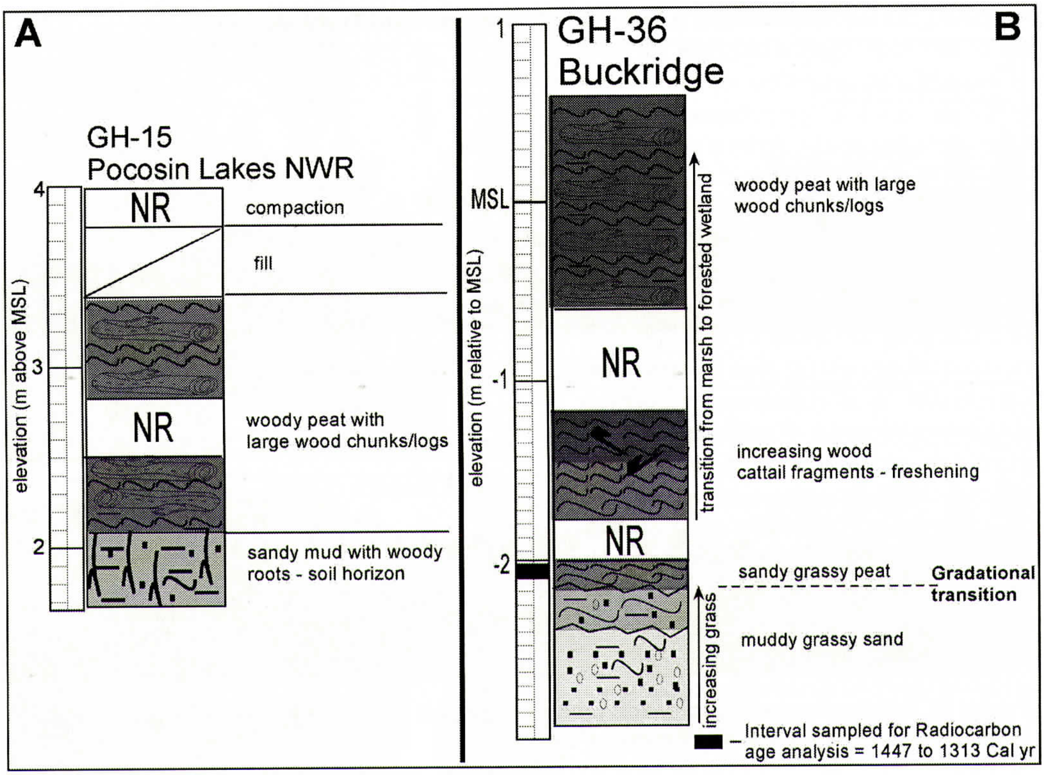

Figure 11. Peat cores from two different areas on the Albemarle/Pamlico Peninsula reflect different successions of environments. In GH-15 (A) from the western portion of the peninsula (see Fig. 1), a rooted soil profile is overlain by woody *in situ* peat indicating a transition from uplands to forested wetlands. However, GH-36 (B) from Buckridge (Fig. 3 for core location) shows no evidence of a rooted soil profile suggesting erosional removal prior to peat accumulation. Note: in order to maximize detail, only portions of each core are shown and elevations of sections are different.

ditions (Sager and Riggs 1998; Riggs et al. 2000; Culver et al. 2008), astronomical tides would have been substantial within the low-lying portions of the Albemarle estuarine system. The Buckridge Coastal Reserve area could have been regularly flooded during high tides resulting in ravinement inland of the location of the present forested wetland peat shoreline (Fig. 12B). Furthermore, astronomical tides could have been amplified by the constricted nature of the Alligator River estuary (Fig. 1). However, plant macrofossil analysis (Table 4) of associated peat indicates that fresh to low-brackish conditions were quickly established and maintained at Buckridge.

As barrier island inlets began to close across the mouth of paleo-Albemarle Sound around 3 ka, tidal amplitudes decreased allowing for marsh development on areas previously inundated during high tide (Fig. 12B). With continued closure of the barrier island system during the last ca. 1 ka (Riggs et al. 2000; Culver et al. 2008), astronomical tidal signals diminished to near zero in the estuaries, as is the case today. Lack of regular flooding tides led to a relative hydrologic stability and allowed for the replacement of marsh with forested wetlands (Fig. 11C) despite slowly increasing sea level (Fig. 12) (Horton et al. 2009; Kemp 2009; Kemp et al. 2009).

Nichols (1989) and Brinson et al. (1995) suggest that a clastic sediment input is required for an estuary-bordering wetland to prograde basinward in response to rising sea level. However,

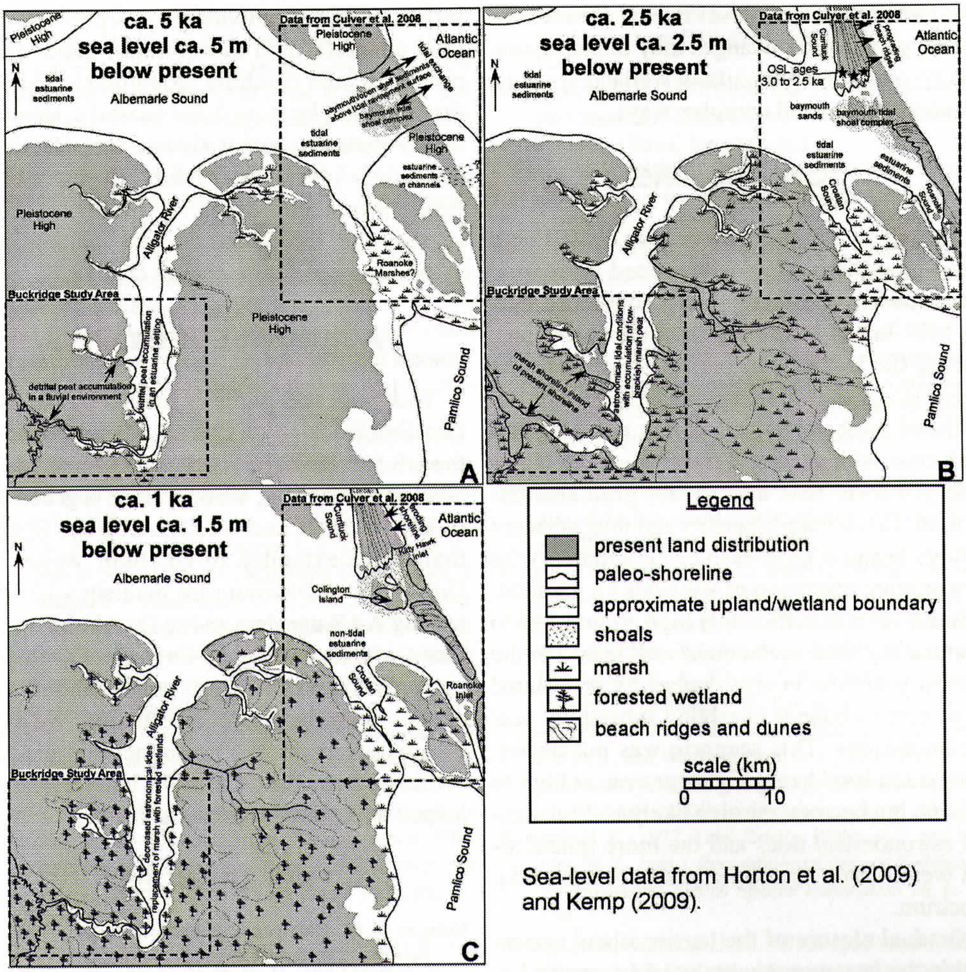

Figure 12. Paleogeographic reconstruction of the Alligator River area based on this study, micropaleontologic and stratigraphic data from the northern Outer Banks area (Culver et al. 2008), and paleo-environmental data from Albemarle Sound (Sager and Riggs 1998) and Croatan Sound (Riggs et al. 2000; Riggs and Ames 2003). Paleo-wetland distribution outside the study areas is based on peat-thickness data (Ingram 1987). Panel A: Albemarle Sound ca. 5 ka was an open embayment with astronomical tidal exchange with the Atlantic Ocean. Detrital peat accumulated at Buckridge and the funnel-shape of the Alligator River probably produced amplified tides. Panel B: By ca. 2.5 ka, sea level was ca. 2.5 m below present and beach ridges prograded across the mouth of Albemarle Sound (Mallinson et al. 2008). Decreased astronomical tidal range promoted marsh progradation over the earlier Holocene tidal ravinement surface in the southern Alligator River area. Panel C: Continued closure of the barrier-islands by ca. 1 ka resulted in much reduced tidal exchange between Albemarle Sound and the Atlantic Ocean and replacement of marsh with forested wetlands in the upper Alligator River area.

at the Buckridge study area, the Holocene sediment record suggests that basinward progradation of wetlands occurred with minimal clastic sediment input. The change in tidal regime from an astronomical tide-dominated system to a system with essentially no astronomical tide is invoked to promote wetland progradation and subsequent peat accumulation to keep pace with rising base level. It is possible that other estuarine systems that have alternated between open marine interchange and enclosure by barrier islands show similar patterns. This finding also demonstrates that sea-level rise and associated coastal geomorphic changes may impact estuarine systems, and particularly wetland systems, in unanticipated and complex ways.

CONCLUSIONS

Buckridge Coastal Reserve is situated on a paleo-topographic high dominated by marine and estuarine sediments that accumulated during MIS 5a and MIS 3 sea-level transgressions. During the last glacial period, these deposits were dissected as drainage patterns became established on the emergent terrain resulting in a landscape with at least 8 m of relief. As Holocene sea level rose and stream gradients decreased, first stream channels and then adjacent valleys became progressively inundated with concomitant migration of wetlands up gradient.

Sea level was sufficiently high around 3 ka to produce bay/tidal ravinement well inland of the present shoreline in areas presently dominated by forested wetlands overlying substantial peat accumulations. This scenario was not dependent on sea level being higher or even as high as present, but because the area likely had substantial astronomical tides and the more inland areas were within the upper portion of the tidal spectrum.

Gradual closure of the barrier island system within the last approximately 2.5 ka resulted in decreasing astronomical tides and freshening of estuarine waters. As a result, marsh prograded seaward over the earlier Holocene bay/tidal ravinement surface. Continued decline of astronomical tides in the Albemarle estuarine system in response to complete closure of all inlets north of Roanoke Island promoted the replacement of fresh to low-brackish marsh with forested wetlands within the last 1 ka. The ultimate result is that the modern forested wetland shoreline, which is presently receding, is well basinward of its ca 3 ka location.

The results from this study illustrate that the response of a coastal region to rising sea level is not simply a matter of progressive inundation of the landscape. Rather, changes in coastal geometry, connectivity between ocean and estuary, and astronomical tidal range can have a major influence on the evolving geomorphology, wetland character and distribution, depositional processes, and resulting stratigraphic patterns. Furthermore, because these coastal systems are highly dynamic, major changes are also likely to continue into the near future. Evidence of paleoenvironmental evolution preserved in the stratigraphic record can provide valuable clues to the nature of future coastal change.

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FLUVIAL TERRACES OF THE LITTLE RIVER VALLEY, ATLANTIC COASTAL PLAIN, NORTH CAROLINA

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ABSTRACT

An optically-stimulated luminescence (OSL) and radiocarbon chronology is presented for fluvial terraces of the Little River, a tributary to the Cape Fear River that drains 880 km² of the Sandhills Province of the upper Coastal Plain of North Carolina. This study differs from previous work in the southeastern Atlantic Coastal Plain in that numerical age estimates are provided for all terraces in the valley of mappable extent by direct dating of fluvial sediments. The Little River valley contains a floodplain and five fluvial terraces with average heights above modern river bed level that range from 3.0 m (T1) to 29.0 m (T5b). Dating indicates the floodplain has a late Holocene (1.3 ± 0.3 ka) to historical age while terraces range in age from 9.9 ± 2.0 ka (T1) to 94.0 ± 15.9 ka (T5b). Age separation of the six fluvial surfaces is corroborated by distinct differences in soil morphology and chemistry. Terrace heights above modern river level and terrace ages indicate a long-term net incision rate of 0.29 mm/yr during the last 100 ka. This rate is nearly an order of magnitude higher than late Pleistocene uplift rates reported for the Cape Fear River valley in the 1980s, based on age estimates for the Wando Formation. However, the 0.29 mm/yr rate is consistent with OSL and radiocarbon dates reported from terraces in the adjacent Pee Dee River valley. Together, these data refine the ages assigned to fluvial facies of the Wando terrace and suggest it is composed of multiple fluvial deposits with a wide range of late Pleistocene ages. The long-term net incision rate of the Little River is consistent with the range of neotectonic uplift rates reported for this region. Scrutiny of the inter-terrace ages indicates that Little River terraces may reflect short-term aggradation in response to periodic climate-mediated increases in sediment supply that is compensated by long-term incision in response to neotectonic uplift.

INTRODUCTION

Fluvial deposits are of great interest to geomorphologists because they provide information about river behavior over long timescales and may contain evidence of past river responses to external forcing mechanisms, such as tectonics, climate change, eustasy, or human impacts (Jacobson et al., 2003; Knighton, 1998). If fluvial deposits can be dated and placed in a chronological framework, then potential drivers and rates of channel change may be evaluated.

The southeastern Atlantic Coastal Plain is composed of unconsolidated, sandy marine sediments, and rivers of this province typically form wide valleys capable of accommodating lateral channel migration while at the same time preserving former deposits. As a result, river valleys in the Coastal Plain commonly contain a relatively well-preserved record of late Quaternary fluvial deposits. This stands in contrast to river valleys in the adjacent Piedmont province, where resistant igneous and metamorphic rocks confine rivers to narrow valleys and inhibit preservation of complete depositional records. Numerous studies have reported age estimates for fluvial terraces in the southeastern Atlantic Coastal Plain (Howard et al., 1993; Leigh et al., 2004; Leigh and Feeney, 1995; Leigh, 2006; Leigh, 2008; Markewich et al., 1987; Markewich et al., 1988; Thom, 1967), but relatively few workers have provided a chronology that encompassed all the terraces present in

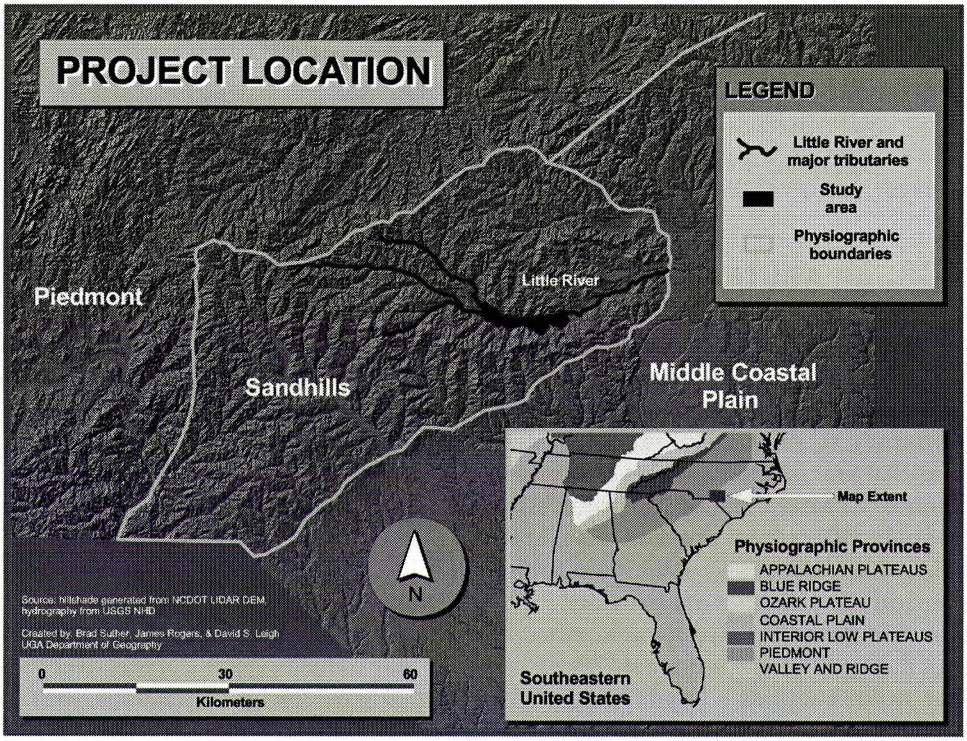


Figure 1. Shaded relief map showing the location of the study area in the Sandhills province of the upper Coastal Plain of North Carolina. Inset shows location of study area within the southeastern United States.

a given valley (Owens, 1989; Soller, 1988). Thus, one objective of this paper is to establish a chronology for fluvial terraces of the Little River, a tributary of the Cape Fear River located in the upper Coastal Plain of North Carolina. This study differs from previous work in the region in that numerical age estimates are provided for all fluvial terraces in the valley by direct dating of fluvial sediment using optically-stimulated luminescence (OSL). Another objective is to present a long-term net incision rate for the Little River for the last 100 ka based on terrace age estimates. The incision rate documented by this study is significant in that it differs from uplift rates reported for terraces of the Cape Fear River (Soller, 1988) and provides insights into incision over late Quaternary timescales in the region.

STUDY AREA

The Little River is a tributary to the Cape Fear River and drains 880 km^2 of the Sandhills province of the upper Coastal Plain of North Carolina (Figure 1). The Sandhills constitute a highly dissected landscape of deeply-weathered, sandy, quartz-rich soils that formed from marine and fluvial deposits of late Cretaceous to Tertiary age (Horton & Zullo, 1991). Surficial geologic mapping indicates that uplands within the Little River basin are comprised of interbedded kaolinitic clays and clayey sands of late Cretaceous age that are overlain in various locations by unconsolidated Tertiary sands and gravels known as the Pinehurst Formation (Conley, 1962). Triassic sandstone and siltstone, felsic tuffs, and late Cretaceous fine sands and clays of marine origin outcrop along the valley sides of the upper Little River and its major tributaries (Conley, 1962). The segment of valley studied is approximately 17 km long and 2 – 3 km wide and is located in the vicinity of the Fort Bragg Military Reservation (Figure 2). The valley contains a floodplain and terraces

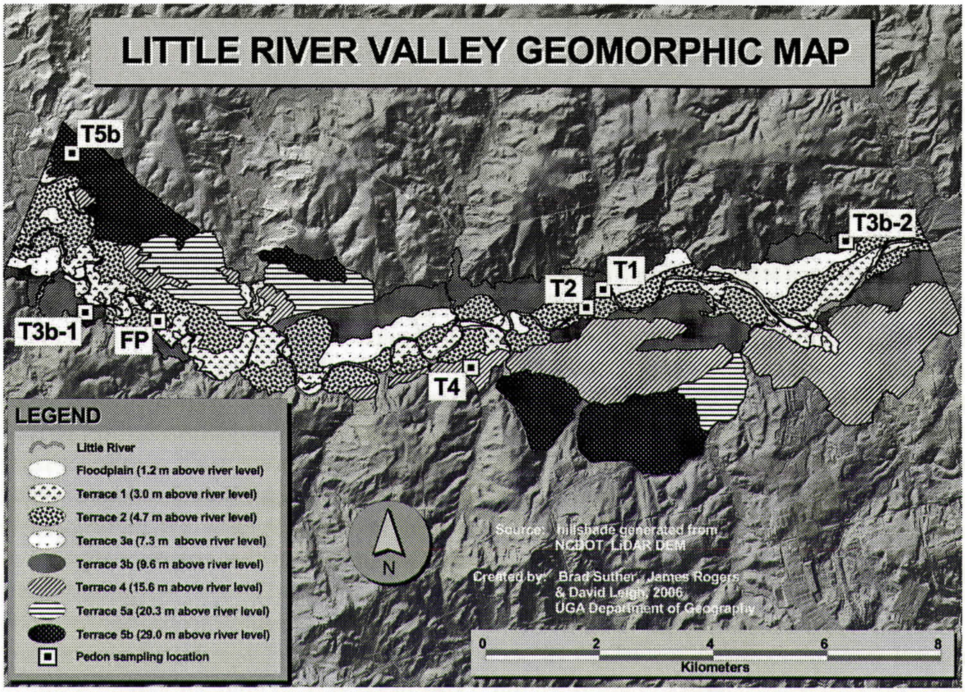


Figure 2. Geomorphic map of the study area. Vertical exaggeration of shaded relief is 3X.

with soils of varying degrees of development that are mapped as ranging from Entisols (Typic Fluvaquents) on the floodplain to Ultisols (Arenic and Typic Hapludults) on higher terraces by the Natural Resource Conservation Service (NRCS) Soil Survey (Hudson, 1984; Wyatt, 1995). The majority of the study area is forested and is dominated by longleaf pine (*Pinus palustris*) savanna, though residential and agricultural landuses are present. The climate is humid subtropical. Average annual precipitation is 117.9 cm, and the average daily temperature is 16.9 °C, with an average January temperature of 6.1 °C and an average July temperature of 27.4 °C (NCDC, 2004). The modern channel has a sandy bed and meandering pattern, with an average slope of 0.00057, a width of 10 to 25 m, and sinuosity of 1.59. The downstream boundary of the study area is located approximately 25 river kilometers upstream of the confluence with the Cape Fear River and is approximately 130 km inland from the present head of the Cape Fear River estuary.

METHODS

The terrace chronology presented in this paper was established to provide a geomorphic and chronologic framework for a soil chronosequence study that evaluated changes in soil properties over time by comparing soils developed on terraces of differing ages (Suther, 2006). Geomorphic mapping was accomplished by interpretation of hillshade and slope derivatives of 2 m and 6 m pixel edge Light Detection and Ranging (LIDAR) digital elevation models provided by the Department of the Army and the North Carolina Department of Transportation (2006), respectively, and US Geological Survey 7.5' topographic quadrangles, followed by field-verification based on visual inspection of the landscape and examination of soil and stratigraphic profiles. Locations of the soil and stratigraphic descriptions and OSL ages presented in this paper were selected to facilitate the soil chronosequence study and were situated to ensure comparability between terrace soils and to reduce the effects of local drainage conditions and topography on soil development.

Because this research was conducted as a component of a soil chronosequence study and a comprehensive characterization of the fluvial history of the Little River was beyond its scope, extensive subsurface investigation to determine the alluvial architecture of the valley was not conducted. Descriptions of soils, sediments, and stratigraphy of terrace deposits were made in backhoe pits at each OSL sampling location using standard NRCS terminology (Soil Survey Division Staff, 1993). Soil descriptions of hand auger samples for a minimum of four additional profiles per terrace were conducted to account for soil variability and are reported elsewhere (Suther, 2006).

OSL Dating

Principles of OSL Dating

OSL dating estimates the time elapsed since sediment was last exposed to sunlight (Stokes and Walling, 2003). After sediment is buried, ionizing radiation from the decay of naturally occurring radionuclides (U, Th, $^{K}40$) in surrounding sediments, and to a lesser extent from cosmic rays, results in the eviction of some electrons in quartz (and other silicate minerals) from their ground state (Aitken, 1998). Evicted electrons become trapped within imperfections in the crystal lattices of these minerals and accumulate over time ($\geq 10^5$ years) until they are exposed to an amount of light or heat energy sufficient to stimulate their return to ground state (Forman et al., 2000). Return to ground state results in emission of excess energy in the form of a photon (luminescence) that is proportional to the total amount of ionizing radiation to which the sample was exposed during burial (Stokes and Walling, 2003). If the irradiation necessary to produce a given luminescence signal can be reconstructed in the laboratory (equivalent dose), and if an annual dose rate from surrounding sediments and cosmic ray contributions can be estimated, a sample age that approximates time of deposition may be calculated by the following equation (modified from Forman et al., 2000)

$$\text{age (ka)} = \text{equivalent dose (Gy)} / \text{dose rate (Gy/ka)} \qquad (1)$$

Where

$$\text{dose rate} = (aD_{\alpha}W + D_{\beta}W + D_{\gamma}W) + D_c \qquad (2)$$

W = water content factor
a = alpha radiation attenuation coefficient
D_{α} = alpha radiation
D_{β} = beta radiation
D_{γ} = gamma radiation
D_c = cosmic radiation

The single aliquot regenerative dose (SAR) protocol (Murray and Wintle, 2000) was used to determine equivalent dose. In this procedure, following measurement of an aliquot's natural luminescence signal, each sample aliquot is subjected to a series of irradiation–pre-heat–stimulation cycles that enable construction of a regenerated growth curve that relates luminescence signal intensity to irradiation dose for each aliquot. The regenerated growth curve is then used to determine an aliquot's equivalent dose. The equivalent dose is the irradiation dose that the growth curve predicts is required to produce the natural luminescence signal observed for a given aliquot. During irradiation–pre-heat–stimulation cycles, the effects of repeat irradiations on luminescence (recuperation effects) are evaluated by determining whether the luminescence intensity generated by a given irradiation dose remains consistent. If necessary, corrections for changes resulting from recuperation effects are incorporated into the equivalent dose estimate.

An underlying assumption of OSL dating is that a sample's luminescence signal is zeroed (bleached) prior to burial (Stokes and Walling, 2003). OSL dating of alluvium may be confounded by partial and/or inhomogeneous bleaching of grains during fluvial transport and deposition (Murray et al., 1995; Olley, 1998, 1999; Stokes and Walling, 2003). However, a number of studies indicate that reliable OSL age estimates can be obtained from fluvial environments (Colls et al., 2001; Leigh et al., 2004; Rittenour et al., 2003, 2005; Srivastava et al., 2001; Stokes et al., 2001; Wallinga et al., 2001). Equivalent dose frequency distributions for Little River samples were inspected for asymmetry indicative of partial bleaching (Olley et al., 1998, 1999). In general, distributions were approximately Gaussian in form, suggesting relatively uniform bleaching of grains. Therefore, the mean equivalent dose was used to calculate OSL age estimates. All ages are reported with two standard deviation error estimates (± 2 sigma).

OSL Dating Procedures

Samples for OSL dating were obtained for at least one terrace per terrace level within the valley (Figure 2). Samples were collected by driving a 4 cm diameter heavy gauge PVC tube laterally into the face of each pit below the depth of bioturbation. C horizon material (unweathered alluvium) was sampled where possible. Sample tubes were sealed immediately after sampling to prevent exposure to light. Sample preparation and handling for OSL dating were carried out in subdued red-light conditions. Five centimeters of sediment were removed from each end of the PVC sample tubes for dose rate estimation. Luminescence measurements were made on samples from the central section of the PVC cylinder that was least likely to have been exposed to sunlight during sampling. All samples were treated with 10% HCl and 30% H_2O_2 to remove carbonates and organic matter. Samples were sieved to extract the 150-170 µm-size fraction. Quartz and feldspar grains were separated by density with Na-polytungstate (ρ=2.58 g cm^{-3}). The quartz fraction was etched using 48% HF for 80 min followed by 36% HCl for 30 min to remove the outer surface affected by alpha radiation. The quartz grains were mounted on stainless steel discs using Silkospray™.

Light stimulation of the quartz was achieved using a RISØ array of blue LEDs centered at 470 nm. Detection optics comprised two Hoya 2.5 mm thick U340 filters and a 3 mm thick Schott GG420 filter coupled to an EMI 9635 QA photomultiplier tube. Measurements were taken with a RISØ TL-DA-15 reader. A 25-mCi $^{90}Sr/^{90}Y$ built-in source was used for sample irradiation.

The single aliquot regenerative dose (SAR) protocol (Murray and Wintle, 2000) used to determine equivalent dose involved a five-point regenerative dose strategy with three dose points to bracket the equivalent dose, a fourth zero dose to test for recuperation effects, and a fifth repeat dose, usually of the smallest change correction incorporated in the SAR protocol. All measurements were made at 125°C for 100 s after a pre-heat to 220°C for 60 s. For all aliquots, the recycling ratio between the first and the fifth point ranged within 0.95-1.05. Data were analyzed using the ANALYST program of Duller (1999).

Equivalent dose measurements were made on single aliquots of 9.6 mm diameter. Typically 15 to 25 aliquots per sample were analyzed. The dose rate calculation relied on the thick source ZnS (Ag) alpha counting technique for elemental concentration of uranium and thorium. Potassium was measured by ICP90, with a detection limit of 0.01%, using the Sodium Peroxide Fusion technique at the SGS Laboratory in Toronto, Canada. The cosmic ray Gamma contribution was assumed to be 150 ± 30 µGy/yr, as recommended for sediments located below an altitude of 1000 m between latitudes 0° - 40° (Prescott and Stephan, 1982). All sample ages were calculated using assumed pore-water content of 15 ± 5% for the in-situ sediment.

Incision Rate Calculations

A net incision rate for the time interval represented by the entire terrace sequence was calculated using the equation

$$IR_{net} = E_{T5b} - E_{mb} / {}_{e}A_{T5b} \quad (3)$$

Where:

IR_{net} = net incision rate
E = elevation
T5b = upper boundary of channel lag sediments of highest terrace
mb = bed of modern river
${}_{e}A$ = estimated age

Net incision rates between intervals of terrace deposition were calculated in a similar manner, by dividing the difference in elevation of the top of channel lag sediments between individual terrace sampling locations by the difference in terrace mean age estimates. The upper boundary of channel lag sediments was chosen as the vertical datum for incision rate calculations because it reflects the natural grade of the river during the time of sediment deposition, it is a laterally traceable surface, and because, where gravels are present, it typically coincides with the depth of refusal when coring. The upper boundary of channel lag in stratigraphic profiles was typically indicated by a pronounced increase, relative to overlying sediments, in the percent weight of >2 mm particles accompanied by a predominantly sand-sized (0.063 – 2.0 mm) matrix. Grain size data were obtained for each profile in vertical increments of 25 cm or finer, respecting soil horizon boundaries, using the hydrometer (Gee and Bauder, 1986) and sieve (Ingram, 1971) methods. The elevation of the upper boundary of channel lag was calculated by subtracting the depth to lag from the ground surface elevation at the sampling location. Ground surface elevations were determined from LIDAR data with a vertical accuracy of ±25 cm (NCDOT, 2006). In nearly all instances, OSL samples were obtained from fluvial sands from stratigraphic positions immediately above or within lag deposits, providing age estimates that approximate lag deposition. Three of the four age estimates obtained for T3b (TU5@190, T32A160, and T32A180) were in close stratigraphic association with channel lag deposits that were encountered at an elevation of 7.1 m and 6.7 m above the modern river bed at the T3b-1 and T3b-2 sampling locations, respectively. The average of the statistical means of these age estimates (59.7 ka), and the average elevation of the top of channel lag at the T3b-1 and T3b-2 sampling locations (6.9 m), were used to approximate T3b age and elevation, respectively, for the purpose of incision rate calculations.

RESULTS

The Little River valley contains a modern floodplain and at least five mappable late Quaternary fluvial terraces (Figure 2). Surface morphology, the soil-stratigraphic profiles associated with OSL sampling locations, and OSL age estimates are described for each terrace below. Morphological attributes of each terrace are summarized in Table 1. OSL age estimates and supporting data are presented in Table 2. Table 3 provides soil profile descriptions of each OSL sampling location.

Terrace Morphology

The floodplain is the alluvial surface being constructed by the modern regime of the Little River and has an average elevation of 1.2 m above the modern river bed. Average height above the modern river bed for the terrace treads ranges from 3.0 m (T1) to 29.0 m (T5b) (Table 1). Terraces 3 and 5 each have two components (denoted a and b), for which the "a" component constitutes a lower and in some instances more eroded surface and the "b" component constitutes a surface that is higher and in some cases more well-preserved. The floodplain and lower terraces are dominated by indeterminant ridge and swale topography from which former river channel patterns are difficult to discern. However, some T1, T2 and T3a surfaces exhibit scrollwork and paleochannel scars clearly indicative of former meandering channels, while others exhibit interwoven ridge and swale topography indicative of former braided channels. Well expressed scrollwork and paleochannel scars are only evident on terraces in the most upstream segment of the study area, while braided patterns are only found in the downstream segment of the study area (Suther, 2006). This variation in terrace morphology may indicate that the study area represents a transition zone along the Little River with respect to past changes in river regime and channel pattern or that individual terraces contain relatively short lived temporal variations in channel pattern. With the exception of one T3b surface that contains indeterminant ridge and swale topography, depositional topography is not apparent on higher terraces (T3b – T5b). In general, terraces exhibit increasing fluvial dissection with increasing height above river level.

Table 1. Elevation, extent, and morphology of terraces within the study area.

Landform	Avg. height above river bed (m)[a]	Proportion of study area (%)[b]	Terrace treads[c] Tread morphology	Degree of dissection
Floodplain	1.2	1.4	flat to indeterminant ridge & swale	none
Terrace 1	3.0	9.1	indeterminant ridge & swale (some surfaces scrolled, others braided)	none
Terrace 2	4.7	19.1	indeterminant ridge & swale (some surfaces scrolled, others braided)	none to slight
Terrace 3a	7.3	6.4	indeterminant ridge & swale (some surfaces possibly braided)	slight
Terrace 3b	9.6	16.0	flat with no depositional topography (one surface contains indeterminant ridge & swale)	slight to moderate
Terrace 4	15.6	23.2	where tread is preserved, flat with no depositional topography (on eroded surfaces only remnant ridges remain)	moderate to high
Terrace 5a	20.3	9.0	where tread is preserved, flat with no depositional topography	moderate
Terrace 5b	29.0	15.9	where tread is preserved, flat with no depositional topography (on eroded surfaces only remnant knobs remain)	moderate to high

[a]Average height of landform surface above modern river bed based on LIDAR DEM data.
[b]Proportion of areal extent of study area represented by each landform.
[c]Qualitative description of terrace tread morphology is based on visual inspection of LIDAR DEM data. Dissection was qualitatively assessed based on the number of intermittent and perennial streams, as mapped on USGS 7.5' topographic quadrangles, that originate on each landform.

Chronology

The modern floodplain contains historical vertical accretion sands and silts deposited since Euro-American settlement of the Little River valley (approximately 200 yr BP) that overlie prehistoric vertical accretion sands and silts of mid- to late- Holocene age. Soils in the floodplain sampling locality consist of Entisols that typically have A-C-Ab-C' profile sequences and contain multiple buried A horizons. The contact between historical and prehistoric sediment at the floodplain sampling location occurs at a depth of 100 cm. The total thickness of floodplain alluvium was not measured and extends below the maximum observed depth of 220 cm. Assuming that deposition of historical sediment initiated 200 yr BP yields an average historical sedimentation rate of 5 mm/yr for the Little River at this site, which is in agreement with historical sedimentation rates documented in other river valleys in the southeastern United States (Leigh and Webb, 2006; Lichtenstein, 2003; Oppenheim, 1996; Costa, 1975; Happ, 1945).

An OSL sample from prehistoric floodplain vertical accretion sediments at a depth of 200 cm within a C''' horizon yielded an age of 1.3 ± 0.3 ka (LR@200). This sample contained concentrations of uranium and thorium that exceed

Table 2. Optically-stimulated luminescence (OSL) dates and supporting data.

Sample	Mean age (ka)	Landform	Depth (cm)	U (ppm)	Th (ppm)	K (%)	Mean Equivalent Dose (Gy)	Dose rate (Gy/ka)
LR@200	1.3±0.3	floodplain	200	23.9±0.7	85.6±30.7	0.96	14.5±1.1	10.9±2.1
31CD475-T2	9.9±2.0	terrace 1	90-120	10.8±1.1	17.9±10.6	0.37	36.3±1.9	3.7±0.7
31CD475-110	17.4±4.2	terrace 2	110	3.9±1.4	14.2±4.9	0.42	37.2±5.0	2.1±0.4
TU5@90	40.0±9.9	terrace 3b-1	90	2.0±0.6	6.9±2.1	0.12	42.4±7.4	1.1±0.2
TU5@190	55.2±15.2	terrace 3b-1	190	1.7±0.3	3.4±1.1	0.20	46.9±11.5	0.9±0.1
T32A160	72.7±13.1	terrace 3b-2	160	1.6±0.3	4.1±0.9	0.18	63.0±9.1	0.9±0.1
T32A180	51.3±12.2	terrace 3b-2	180	2.60±1.0	9.90±3.5	0.21	74.1±8.0	1.4±0.3
T4-161	74.6±10.4	terrace 4	161	4.0±0.8	12.6±2.6	0.30	145.6±6.4	2.0±0.3
T5-135	94.0±15.9	terrace 5b	135	3.8±0.7	5.4±2.8	0.10	123.1±6.6	1.3±0.2

Notes: Sample water contents were assumed to be 15±5% by weight. Cosmic ray contribution was assumed to be 150±30 μGy per annum.

levels typically found in natural systems, which resulted in an unusually high dose rate estimate (Table 2), and suggests possible contamination of the sediments. Because overestimation of dose rate may result in underestimation of the true age of the sample, the authors regard this age with caution. However, age estimates of deposits associated with the next highest geomorphic surface in the terrace sequence, T1, which include a 9.9 ± 2.0 ka OSL date (this paper) and a 10.24 ± 0.03 ka radiocarbon date (Goman and Leigh, 2004, discussed below), coupled with a lack of pedogenic development in floodplain sediments (Table 3, this paper; Suther, 2006), provide context that suggests the prehistoric floodplain is Holocene in age, with a mid- to late- Holocene age being most likely. Assuming the floodplain profile was not truncated by erosion prior to the deposition of historical sediment, an age of 1.3 ± 0.3 ka would suggest that floodplain sedimentation at the sampling location proceeded at an average rate of 0.6 – 1.0 mm/yr between about 1 ka and the initiation of historical sedimentation. Though this value is not unreasonable, it is on the upper end of typical prehistoric sedimentation rates observed in river valleys of the southeastern United States (Leigh and Webb, 2006; Lichtenstein, 2003; Leigh, 1996).

Terrace 1 deposits exhibit roughly normal grading and consist of lateral accretion medium and coarse gravels and coarse sands that fine upward to vertical accretion sands. Soils in the T1 sampling locality consist of Inceptisols containing eluviated (E) horizons above incipient B horizons that lack clay coatings but have weak, medium subangular blocky structure and show signs of iron oxide accumulation. An early Holocene to terminal Pleistocene age (9.9 ± 2.0 ka, 31CD475-T2) is indicated for T1 deposits by an OSL sample taken from sandy lateral accretion deposits in a C1 horizon at a depth of 90 – 120 cm near the location of the T1 backhoe pit (description provided in Table 3). The calculated optical age estimate is in agreement with independent ^{14}C dates of peat sampled from the thalweg and point bar of an abandoned paleochannel on a T1 unit about 2.2 km west of the T1 OSL sampling location (Goman and Leigh, 2004). The basal age of peat was found to be 10.24 ± 0.03 ka (Beta-151615) in the thalweg of the paleochannel and 9.23 ± 0.19 ka (Beta-152729) on the point bar. Goman and Leigh note the discrepancy in age between the thalweg and point bar basal peats likely reflects the control of depositional setting on peat accumulation, with peat accumulating first in the low elevation thalweg and later above the point bar, which is at a slightly higher elevation.

Terrace 2 deposits are normally graded and consist of lateral accretion deposits of medium gravels and coarse sands that fine upward to vertical accretion sands. Soils in the T2 sampling locality are predominately Ultisols that contain eluviated (E) horizons underlain by argillic (Bt) horizons with moderate medium sub-

Table 3. Soil profiles at OSL sampling locations. These profiles are representative of those characterized by Suther (2006) and are typical of soils found on the middle part of the upper surface of each respective terrace in the most well-drained and least eroded landscape position.

Landform	Horizon	Depth (cm)	Moist matrix color[a]	(>2 mm)[b] (%)	Sand[c] (%)	Silt[c] (%)	Clay[c] (%)	Structure[d]
FP	A	0-7	10YR 4/2	0.0	85.9	9.6	4.5	wk med gr
	C1	7-24	10YR 6/4	0.0	92.1	5.1	2.7	wk med gr
	C2	24-37	10YR 5/4	0.0	84.6	12.1	3.2	wk med gr
	A'b	37-47	10YR 4.5/4	0.0	92.0	5.2	2.8	wk med gr
	C'1	47-58	10YR 7/3	0.0	97.5	1.8	0.7	sg
	C'2	58-78	10YR 6/6	0.0	95.0	3.3	1.7	wk fn gr
	C'3	78-100	10YR 5.5/6	0.1	91.9	5.5	2.6	wk med gr
	A"b	100-114	10YR 5/4	0.1	94.0	4.2	1.8	wk fn gr
	C"1	114-130	2.5Y 6/4	0.0	96.7	2.1	1.3	wk fn gr
	C"2	130-170	10YR 5/4	0.0	95.3	3.2	1.5	wk fn gr
	A'"b	170-187	10YR 4/3	0.0	84.7	12.0	3.3	mod fn gr
	C'"	187-220+	10YR 4/4	0.0	78.5	15.1	6.4	mod fn gr
T1	Ap	0-8	10YR 3/2	0.1	94.1	3.9	2.0	wk fn gr
	E1	8-19	2.5Y 5/4	0.1	93.1	4.9	2.0	wk fn gr
	E2	19-62	10YR 6/4	0.1	92.3	6.2	1.5	wk fn gr
	Bw	62-81	10YR 6/6	1.3	91.2	6.9	2.0	wk med sbk
	BC	81-88	10YR 5/8	5.4	94.4	4.1	1.5	wk med gr
	C1	88-109	10YR 6/6	30.2	96.7	2.6	0.7	sg
	C2	109-128	10YR 6/4	15.4	97.9	1.3	0.8	sg
	C3	128-166	10YR 7/6	21.5	97.9	0.9	1.2	sg
	C4	166-211+	10YR 7/6	45.3	97.7	1.3	1.0	sg
T2	A	0-10	10YR 4/3	0.2	85.5	11.4	3.0	wk fn gr
	EA	10-20	2.5Y 5/4	0.1	87.7	9.8	2.5	wk fn gr
	E	20-60	2.5Y 6/4	1.1	83.7	12.2	4.1	wk fn gr
	Bt	60-100	7.5YR 5/7	1.6	77.9	10.7	11.4	mod med sbk
	C	100-125	10YR 6/6	8.3	96.1	1.1	2.8	sg
	Csm	125-140	5YR 5/8	33.0	93.8	1.2	5.0	sg - cemented
	C'	140-160+	2.5Y 6/3	49.2	94.1	2.0	3.9	sg
T3b[e] (site 1)	Ap	0-18	2.5Y 3/1	0.4	89.3	8.3	2.5	wk med gr
	E1	18-39	2.5Y 6/4	0.6	90.8	7.3	1.9	wk fn sbk
	E2	39-67	2.5Y 7/4	3.8	91.9	6.4	1.7	wk fn sbk
	Bt	67-98	10YR 5/6	6.6	79.5	4.7	15.8	mod fn sbk
	C	98-103	2.5Y 7/3 & 2.5Y 7/4	3.3	96.0	2.2	1.7	sg
	Btb1	103-116	10YR 5/6	1.7	80.9	6.5	12.6	mod fn sbk
	Btb2	116-140	10YR 7/4 (dep)	2.3	68.6	13.6	17.8	mod fn sbk
	Btb3	140-166	10YR 7/4 (con, dep)	4.0	67.4	19.2	13.4	mod fn sbk
	C'	166-210	2.5Y 7/4 (con, dep)	8.0	98.3	0.2	1.4	sg
	2C	210-230+	2.5Y 7/1 & 5YR 6/4	0.0	11.1	46.6	42.2	ma

[a]Abbreviations: (con) = redox concentrations present, (dep) = redox depletions present.

[b]Percentages based on dry wt. of >2 mm fraction.

[c]Percentages based on dry wt. of <2 mm fraction.

[d]Abbreviations: wk = weak, mod = moderate, fn = fine, med = medium, cs = coarse, gr = granular, sbk = subangular blocky, abk = angular blocky, sg = single grained, ma = massive.

[e]OSL sample TU5@90 was obtained from a depth of 90 cm in this C' horizon about 1m away from the described profile, where the C' horizon occurs at a slightly shallower depth. Lower boundary of Little River alluvium is 210 cm.

Table 3, continued. Soil profiles at OSL sampling locations. These profiles are representative of those characterized by Suther (2006) and are typical of soils found on the middle part of the upper surface of each respective terrace in the most well-drained and least eroded landscape position.

Landform	Horizon	Depth (cm)	Moist matrix color[a]	(>2 mm)[b] (%)	Sand[c] (%)	Silt[c] (%)	Clay[c] (%)	Structure[d]
T3b (site 2)	Ap	0-6	10YR 3/2	0.1	89.5	9.0	1.5	wk fn gr
	E1	6-42	2.5Y 5/4	0.2	89.6	8.6	1.8	sg
	E2	42-73	2.5Y 6/4	0.3	88.3	9.4	2.3	sg
	Bt1	73-100	10YR 5/8 (con)	0.5	88.5	5.4	6.1	wk med sbk
	Bt2	100-147	10YR 5/8	2.4	95.8	1.5	2.7	wk fn sbk
	C & Bt	147-170	2.5Y 7/3 (C) 10YR 5/8 (Bt)	2.9 (C) 1.4 (Bt)	99.1 (C) 96.8 (Bt)	0.5 (C) 0.9 (Bt)	0.4 (C) 2.3 (Bt)	sg (C) wk med sbk (Bt)
	C1	170-190	10YR 7/8 (con, dep)	5.5	98.6	0.3	1.1	sg
	C2	190-227	10YR 7/6 (con,dep)	7.3	98.5	0.5	1.0	sg
	C3	227-235+	10YR 7/1	14.4	99.3	0.2	0.5	sg
T4[e]	Ap	0-11	2.5Y 3/1	0.1	87.7	10.8	1.5	wk fn gr
	E	11-54	2.5Y 6/4	0.5	87.0	11.4	1.6	wk fn gr
	Bt	54-90	10YR 5/8	1.9	85.6	9.1	5.3	wk med sbk
	BE	90-130	10YR 6/6 & 10YR 6/8	4.2	90.8	7.4	1.8	wk med gr
	E'	130-166	10YR 7/6 & 2.5Y 7/4	9.8	90.4	8.0	1.6	sg
	B't1	166-204	7.5YR 5/8 (dep)	27.1	84.6	5.7	9.7	wk med sbk
	B't2	204-252	7.5YR 5/8 (con)	6.2	82.8	4.7	12.5	wk to mod med sbk
	B't3	252-287	10YR 7/6 (con, dep)	2.4	74.2	6.5	19.3	mod med sbk
	B't4	287-307	10YR 6/8	2.1	80.3	4.3	15.4	mod med sbk
	B't5	307-347	10YR 7/1 & 10YR 7/2	0.2	51.4	20.5	28.1	no data
T5b[f]	A	0-15	2.5Y 3/2	0.0	90.0	7.5	2.5	wk fn gr
	Ap	15-40	2.5Y 4/2	0.1	89.0	9.3	1.8	wk med gr
	E	40-63	2.5Y 5/4	0.0	84.0	13.3	2.7	wk fn gr
	Bt	63-87	10YR 5/8	0.2	83.1	10.7	6.3	wk med sbk
	E'1	87-113	10YR 6/6 (con, dep)	0.2	87.8	9.7	2.5	wk med sbk
	E'2	113-134	10YR 6/6 (con, dep)	0.3	88.8	8.7	2.5	wk fn gr to sg
	B't	134-145	10YR 5/6 (con)	0.4	82.4	11.6	6.0	wk med sbk
	Btx1	145-180	10YR 5/8, 2.5Y 6/3, 2.5Y 6/2 (con, dep)	0.7	77.4	14.6	8.0	wk med sbk
	Btx2	180-256	2.5YR 4/8, 10YR 5/8, 10YR 6.5/1 (con, dep)	1.3	70.3	8.1	21.6	wk med sbk to ma
	Btx3	256-274	2.5YR 4/8, 5YR 5/8, 10YR 6/2, 10YR 6/1 (con, dep)	6.9	72.0	10.5	17.5	wk med sbk to ma
	Btx4	274-288	7.5YR 6/6 (con)	7.6	75.2	11.8	13.0	wk med sbk to ma
	Btx5	288-300	2.5Y 4/8, 7.5YR 6/8, 10YR 7/1, 10YR 7/2 (con, dep)	11.5	68.6	11.5	19.9	wk med sbk to ma
	Btx6	300-318	10YR 7/2 (con)	20.5	61.1	12.5	26.4	wk cs abk to ma
		318-360	10YR 7/2 (con)	5.4	51.6	19.4	29.0	

[a]Abbreviations: (con) = redox concentrations present, (dep) = redox depletions present.

[b]Percentages based on dry wt. of >2 mm fraction.

[c]Percentages based on dry wt. of <2 mm fraction.

[d]Abbreviations: wk = weak, mod = moderate, fn = fine, med = medium, cs = coarse, gr = granular, sbk = subangular blocky, abk = angular blocky, sg = single grained, ma = massive.

[e]Description is composited from two pedons for illustrative purposes. Ap-B't2 horizons are taken from T4, pedon 1, and B't3-B't4 horizons are taken from T4, pedon 2 (Suther, 2006). Sampling limitations at pedon 1 prevented description of the entire profile. Lower boundary of alluvium is 307 cm.

[f]Lower boundary of alluvium is at 318 cm within the Btx6 horizon.

angular blocky structure and sand grains that are coated and bridged with clay. An age of 17.4 ± 4.2 ka (31CD475-110) is indicated for T2 by an OSL sample taken from sandy lateral accretion deposits in a C1 horizon at a depth of 110 cm. The complete alluvial thickness was not measured for T1 and T2 deposits, but based on descriptions that extend well into bedload sediment, it is greater than 211 cm for T1 and 160 cm for T2.

Deposits associated with two separate treads of T3b (T3b-1 and T3b-2) were dated. Ultisols have formed in sediments at both localities that typically have A-E-Bt1-Bt2-C profile sequences, with a third Bt or C & Bt horizon sometimes present. At the T3b-1 sampling locality, two allostratigraphic units are present, a lower unit (210 to 103 cm), interpreted as a truncated paleosol formed in lateral accretion sands; and an upper unit (103 to 0 cm), composed of a sand splay deposit overlain by lateral accretion sands that fine upward to vertical accretion sands and silts. The total thickness of alluvium at the T3b-1 locality is 210 cm. An OSL date of 40.0 ± 9.9 ka (TU5@90) was obtained from C horizon sand at a 90 cm depth associated with the sand splay deposit in the upper allounit. A date of 55.2 ± 15.2 ka (TU5@190) was obtained from sandy lateral accretion sediments in a C' horizon near the base of the lower allounit at a depth of 190 cm. The authors consider these ages to be in good agreement. Although in theory deposits at 90 cm and 190 cm could be the same age due to overlap in the 2-σ error of their age estimates, the allostratigraphy at location T3b-1 clearly indicates that sediments at 190 cm are older than those at 90 cm.

Terrace 3b-2 deposits exhibit normal grading and consist of lateral accretion coarse sands and fine gravels that fine upward into vertical accretion sands and silts. The total thickness of alluvium at the T3b-2 locality was not determined, but based on a description that extends well into bedload sediment, it is greater than 235 cm. At the T3b-2 locality, an OSL date of 72.7 ± 13.1 ka (T32A160) was obtained from sandy lateral accretion deposits in the C component of a C & Bt horizon at a depth of 160 cm, while a date of 51.3 ± 12.2 ka (T32A180) was obtained from sandy, lateral accretion deposits lower in the profile in a C1 horizon at 180 cm. The statistical means of these age estimates are inconsistent with their stratigraphic position. However, given the overlap in the 2-σ error of these estimates, it is possible that sediments at 160 cm are the same age as or slightly younger than those at 180 cm, as their stratigraphic positions indicate. Therefore, the authors regard them to be in rough agreement. Relative consistency in age estimates exists both within the individual T3b-1 and T3b-2 stratigraphic profiles and between the separate terrace units. This consistency lends support to the reliability of OSL age estimates presented in this study. Because T3a deposits tended to be less well preserved than those of T3b, they were not selected for dating.

Terrace 4 and T5b deposits consist of lateral accretion medium to coarse gravels and coarse sands that fine upward to vertical accretion sands and silts at the sampling localities. Soils on T4 and T5b are deeply weathered Ultisols and exhibit bisequal soil profiles with multiple argillic (Bt) horizons that are present to the base of and extend below the contact with underlying alluvial fill and/or Coastal Plain sediments that predate the deposition of T4 and T5b (Table 3; Suther, 2006). The total thickness of alluvium at the sampling locality is 307 cm for T4 and 318 cm for T5b. Because unweathered C horizon material was not available for T4 or T5b, OSL ages for both terraces were obtained from E horizons that contained about 90% sand by weight. An OSL sample taken from sandy, lateral accretion sediments at a depth of 161 cm in the E' horizon indicated an age of 74.6 ± 10.4 ka (T4-161) for T4 (Table 2). Although there is overlap in the 2-σ error in the age estimates for T4 and both dated units of T3b (samples TU5@190 and T32A160, Table 2), T4 is clearly older based on its greater height above river level (Table 1) and the degree of pedogenic development characteristic of T4 deposits (Table 3, this paper; Suther, 2006). Terrace 5b is estimated to be 94.0 ± 15.9 ka (T5-135) based on an OSL age obtained from sandy, vertical accretion sediments at a depth of 135 cm in the E'2 horizon. The 2-σ error of this date falls within the range of error of the T4 age estimate and two dates obtained from T3b (TU5@190 and T32A160, Table 2). However, the height of T5b above the modern river (Table 1) clearly indicates it is older than both T4 and T3b. Because T5a deposits were typically less well preserved than those of T5b, they were not selected for dating.

In summary, OSL age estimates indicate a mid- to late-Holocene age for prehistoric sediments of the active floodplain (≥1.3 ± 0.3 ka), an early Holocene age for T1 (9.9 ± 2.0 ka), and a late Wisconsin to terminal Pleistocene age for T2 (17.4 ± 4.2 ka). Age estimates for T3b sediments range from 40.0 ± 9.9 ka to 72.7 ± 13.1 ka (30.1 to 85.8 ka, considering the full breadth of 2-σ error of all T3b ages), indicating T3b may have been active from as early as OIS 5a through late OIS 3. Age estimates for T4 and T5b are 74.6 ± 10.4 ka (OIS 5a through OIS 4) and 94.0 ± 15.9 ka (OIS 5d – 5a), respectively. Although overlap exists in the 2-σ error of T3b, T4, and T5b ages, height above modern river level indicates T3b is the youngest of these terraces while T5b is the oldest, as the statistical means of age estimates suggest. Age separation of the six fluvial surfaces is also corroborated by a relative increase in pedogenic development of soils across the terrace sequence. From lower to higher surfaces, soils show a progressive increase in Bt horizon thickness, subsoil clay content, redness, and dithionite-extractable iron; a progressive decrease in the ratio of bases (CaO, Na_2O, K_2O, MgO) to alumina (Al_2O_3) and other resistant oxides in the bulk chemistry of the whole soil (<2 mm) fraction; and an increasingly mature clay fraction mineral assemblage (Suther, 2006). Despite the low precision and resolution of age estimates for older terraces, these data are significant in that they clearly indicate that the five mappable fluvial terraces within the Little River valley were deposited within the last 100 ka.

Incision Rates

Little River terraces provide a record of fluvial incision since the time of their deposition, and terrace age estimates allow the calculation of net incision rates during the late Pleistocene and Holocene (Table 4). An age of 94.0 ± 15.9 ka for T5b indicates that the long-term net incision rate for the Little River in the vicinity of the study area has been about 0.29 mm/yr (0.25 – 0.35 mm/yr). If the full breadth of 2-σ error in OSL age estimates is considered, then it is possible that incision has been relatively constant during the last 100 ka. However, incision rates based on the statistical mean of OSL age estimates vary during this period and range from 0.65 mm/yr between the deposition of T5b and T4 (94.0 to 74.6 ka) to 0.07 mm/yr between the deposition of T3b and T2 (59.7 to 17.4 ka). A radiocarbon date from a depth of 246 cm obtained from basal peat overlying lag gravel in the thalweg of an abandoned paleochannel on T1 indicates that the Little River incised to its present bed elevation before 10.24 ± 0.03 ka (Goman and Leigh, 2004) and that the river has migrated laterally since that time.

It is important to note that OSL samples were obtained from soil chronosequence sampling locations rather than from locations in the valley spatially representative of the entire interval of aggradation recorded by each terrace. Age estimates presented here have neither the spatial nor temporal resolution sufficient to precisely bracket intervals of aggradation and abandonment of former floodplains. As a result, the incision rates reported above should be regarded as representing the long-term *net* incision that has resulted from the combined processes of incision and aggradation that have operated since the time of active fluvial sedimentation documented by the OSL date.

DISCUSSION

Comparisons with Previous Studies

A net incision rate of 0.29 mm/yr (0.25 – 0.35 mm/yr) for the Little River during the past 94.0 ± 15.9 ka is significant because it is nearly an order of magnitude higher than the uplift rate of 0.064 mm/yr (0.21 ft/1000 yr) reported by Soller (1988, p. A49) for the past 100 ka for the northwestern portion of the lower Cape Fear valley. Soller (1988) estimated uplift rates for the lower Cape Fear valley using a best-fit river

Table 4. Net incision rates for the Little River.

Landform	Incision[a] (m)	Time interval[b] (ka)	O isotope stage	Net incision rate (mm/yr)[c]
T1 to modern river bed	0.0 - 0.0	10.2 to 0.0	early to late 1	0.00
T2 to T1	4.0 - 0.0 = 4.0	17.4 to 10.2	2 to early 1	0.56
T3b[d] to T2	6.9 - 4.0 = 2.9	59.7 to 17.4	late 4 to 2	0.07
T4 to T3b[d]	14.4 - 6.9 = 7.5	74.6 to 59.7	early to late 4	0.50
T5b to T4	27.0 - 14.4 = 12.6	94.0 to 74.6	5b to late 4	0.65
T5b to modern river bed	27.0 - 0.0 = 27.0	94.0 to 0.0	5b to late 1	0.29

[a]Thickness of vertical incision based on difference in elevation above river bed of upper boundary of channel lag sediments between terraces. T3b lag elevation is an average of lag depth in the T3b-1 and T3b-2 profiles.
[b]Based on means of OSL dates (T2-T5b) and radiocarbon date (T1, Goman & Leigh, 2004).
[c]Rates were calculated by dividing incision thickness by the difference in terrace mean age estimates.
[d]Age estimate for T3b is an average of the statistical means of age estimates from OSL samples TU5@90, T32A160, and T32A180, which were in close stratigraphic association with lag sediments.

terrace longitudinal profiling technique and attributed incision of the Cape Fear River over the late Pliocene and Pleistocene to tectonic uplift associated with a series of local flexures that were superimposed on a gentle, persistent, regional uplift of the Cape Fear arch. Although the Little River incision rate and Soller's uplift rate were determined by different techniques, the discrepancy between them is nonetheless surprising given that the Little River is a major tributary to the Cape Fear River located on the southwestern limb of the Cape Fear arch in close proximity (~ 45 km) to Soller's study area. Because net incision rates calculated for most intervals between the deposition of dated Little River terrace sediments (Table 4) are also substantially higher than Soller's estimate, it seems unlikely this discrepancy results from underestimation of the age of T5b.

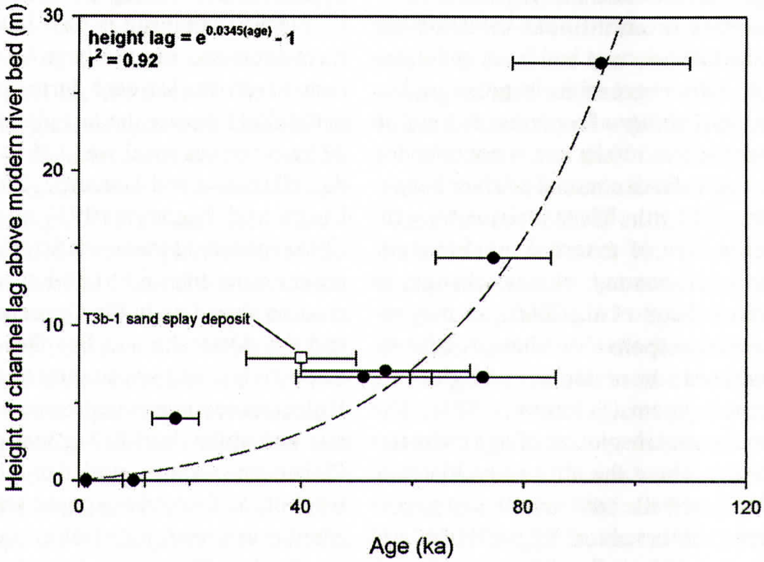
If uplift has been greater near the Piedmont, as Soller (1988) contends, it is possible the Little River, located in closer proximity to the Piedmont than Soller's study area, experienced more rapid local uplift than the lower Cape Fear River and responded with more rapid incision. Markewich (1985), working along the upper Cape Fear River near its confluence with the Little River, observed that the modern upper Cape Fear channel is incised 13 to 28 m beneath its lowest terrace. She suggested incision was in response to late Pleistocene uplift of the Sandhills upland north and west of Fayetteville, North Carolina (which includes the Little River catchment), along a flexure or zone of faulting that extends from the Rockfish Creek – Cape Fear River confluence northeast to Smithfield, North Carolina. Markewich estimated a 60 – 200 ka age for the lowest terrace of the upper Cape Fear River based on soil properties and two pieces of wood that yielded radiocarbon ages of >40 ka. Although incision rates calculated by the authors using a 200 ka age for this terrace (0.07 – 0.14 mm/yr) are lower than the Little River rate, rates based on the 60 ka minimum age for this terrace (0.22 – 0.47 mm/yr) are comparable to those of the Little River and illustrate how possible local variation in neotectonic uplift could account for the difference between the Little and Cape Fear River rates.

A more likely explanation for the discrepancy between Soller's uplift rate and the Little River incision rates is that the estimate of 100 ka for the Wando terrace, used to calculate Soller's rate for the Cape Fear valley during the late Pleistocene, is an overestimate. The Wando terrace, which also is present in the Pee Dee River valley, was not directly dated but assigned an age of 90 ka by Owens (1989). This age was determined by correlation to the Wando Formation, a marine deposit near Charleston, South Carolina, whose age is estimated by amino acid racemization and uranium-series dating of corals to be between 87 – 126 ka (McCartan et al., 1982; McCartan et al., 1980). However, Leigh et al. (2004, pp. 73 – 74) reported several OSL and radiocarbon dates that indicate the Wando terrace in the Pee Dee River valley was active ca. 13 – 22 ka. A younger age for this terrace is also reported by Thom (1967), who indicated it was active shortly prior to 17,000 and 36,000 ^{14}C yr BP based on radiocarbon dates from samples beneath dunes. Furthermore, geomorphic mapping by Leigh et al. (2004, p. 74) indicates that the Wando terrace in the Pee Dee valley is actually comprised of multiple terrace levels with different ages, based on cross-cutting relationships and differences in terrace elevations and surface morphologies. The Little River is a tributary to the Cape Fear, and some differences in drainage basin characteristics and external and internal drivers of aggradation and incision certainly exist between the two systems. However, it seems unreasonable that the valley of a major tributary such as the Little River would contain five late Quaternary fluvial terraces while the Cape Fear valley contains only one. Given this evidence and assuming that the Wando terraces in the Pee Dee and Cape Fear valleys are indeed correlative with one another, it is probable that the Wando terrace in the Cape Fear valley also contains deposits considerably younger than 90 – 100 ka. Overestimation of the age of the Wando terrace would result in an underestimation of the rate of incision since its deposition and could account for the discrepancy between the incision rates of the Little River and the Cape Fear River. Although data are insufficient to resolve correlations between Little River terraces and late Pleistocene terraces of the Cape Fear and Pee Dee Rivers, the findings of this study and Leigh et al. (2004) refine age estimates for the fluvial facies of the Wando Formation of Owens (1989) and indicate it is composed of multiple fluvial deposits with a range of late Pleistocene ages.

Mills (2000) compiled age-height data from 16 studies of terraces and cave sediments along rivers of the eastern United States ranging in age from Holocene to >10 Ma. After applying the scaling function of Gardner et al. (1987) in an effort to correct for the effect of measured time interval on calculated incision rate, he observed a 19 fold decrease in incision rate for a 10^3 increase in measured time interval and concluded that incision may have increased during the late Cenozoic. The Little River incision rate is within the magnitude of incision rates calculated by the authors from terrace height-age pairs compiled by Mills from dated terraces <100 ka in the Appalachian and Coastal Plain Provinces (0.1 – 1.0 mm/yr) and suggests the Little River has experienced incision rates similar to other rivers of the eastern United States during this time period. However, incision rates calculated from ≥100 ka - >10 Ma deposits, which typically range from 0.01 – 0.10 mm/yr, are about an order of magnitude lower than the Little River rate (Mills, 2000). Because Mills summarized incision from rivers across the eastern United States over the past 10 Ma, possible local and temporal variations in tectonics and responses to past climate change and eustasy prevent definitive conclusions about the relationships between Little River incision and that of other rivers. However, considered in the context of the data compiled by Mills, the record of Little River incision may lend support to the contention that rivers in the eastern United States show a general trend of increasing incision through the late Cenozoic.

Incision rates determined from fluvial terraces preserved in unglaciated catchments of northwestern Europe allow comparison of the long-term net incision of the Little River with that of rivers in another temperate, passive mar-

Figure 3. Height of upper boundary of channel lag above modern river bed at each OSL sampling location versus OSL age estimate (n=8). Floodplain and terrace 1 lag sediments typically occur at the approximate elevation of the modern river bed and are assigned a height of zero. Error bars give 2-sigma error of OSL age estimates. OSL sample TU5@90, associated with a sand splay deposit, was not included in the regression because it postdates lag sediments but is shown on the figure for illustrative purposes.



gin setting that experienced substantial climatic change during the Quaternary. In general, the late Pleistocene long-term net incision rate of the Little River is nearly an order of magnitude greater than late Pliocene and Quaternary rates reported for rivers of northwestern Europe and southern England that lie outside the limits of the Pleistocene glaciations. Antoine et al. (2000) report net incision rates of 0.05 – 0.06 mm/yr during the past 1 Ma for the middle courses of the Seine River and Somme River valleys (northwestern France), while van den Berg (1994) reports a similar rate of 0.061 mm /yr over the past 2.4 Ma for the Maas River (The Netherlands). A comparable rate of 0.07 mm/yr is reported by Maddy (1997) for the upper course of the Thames River (southern England) during the past 1.8 Ma, although rates were slightly higher (0.13 – 0.14 mm/yr) during the past 440 ka for the Lower Thames (Maddy et al, 2000). According to these authors, terrace deposits reflect periods of aggradation that occurred during cold-climate, glacial or stadial conditions in response to high sediment yield under periglacial landscape conditions, whereas downcutting and terracing occurred rapidly in response to neotectonic uplift during interglacial or early glacial, transitional periods when climate was warmer and the sediment:discharge ratio was favorable to incision. Although the long-term net incision rate of the Little River is considerably greater than rates reported for northwestern European rivers, evaluation of Little River inter-terrace incision rates indicates that similar interactions between uplift and climate-mediated sediment supply may have contributed to the formation of the Little River terrace sequence (see below).

Possible variation in Little River incision during the last 100 ka

If the 2-σ errors of age estimates of Little River terraces are considered, it is possible that the Little River incised at a relatively constant rate during the last 100 ka. However, incision rates based on mean ages indicate that incision has varied during this time (Table 4). Figure 3 depicts height above the modern river bed versus age for channel lag sediments at each OSL sampling location. Although sample size is too small to permit sound statistical analysis of this data set, nonlinear regression employing an exponential growth function was performed simply to illustrate the nonlinear relationship between height above river bed level and mean terrace age. This regression is presented to highlight the *possibility* of a nonlinear trend in incision over the last 100 ka and is not intended to serve as a quantitative model of river behavior. Variations in Little River incision may reflect the influence of external mechanisms, such as tectonics, eustasy, climate change, or glacioistostasy (Scott et al., 2009), or may reflect a "complex response" to changes with respect to aggradation or incision originating within a river system (Schumm, 1973). The temporal and spatial resolution of age estimates and information about the alluvial architecture of the Little River valley are insufficient to precisely bracket intervals of aggradation and abandonment of former floodplains. Although this prevents definitive conclusions about incision during the late Pleistocene, the Little River data allow for some interpretations and speculations, and possible forcing mechanisms are evaluated below.

The southeastern Atlantic Coastal Plain is located on a passive margin and is characterized by relatively slow rates of neotectonic uplift ranging from 0.01 to 1.8 mm/yr (Cronin, 1981; Marple and Talwani, 2000; Soller, 1988). The long-term net incision rate of the Little River falls within this range, and net incision over the past 100 ka may be in part a response to neotectonic uplift. Soller (1988) concluded that the spatial arrangement of terraces of the Cape Fear River resulted from tectonic uplift associated with a series of local flexures that were superimposed on a gentle, persistent, regional uplift of the Cape Fear arch. Soller did not report variation in uplift within the time interval represented by Little River terraces (late Pleistocene and Holocene). The Little River is a tributary to the Cape Fear located on the southwestern limb of the Cape Fear arch, and it seems reasonable that neotectonic activity played some role in the evolution of its present valley. However, the authors are unaware of any evidence that indicates the vertical or planimetric distribution of Little River terraces can be solely explained by incision in response to variations in uplift of the Cape Fear arch during the last 100 ka.

Eustasy can result in base level changes that force incision, and the base level of the Cape Fear River was lowered during glacial periods, particularly during the last glacial maximum at 22 ka when sea level was 125 m lower than today (Balsillie and Donahue, 2004). However, Leigh and Feeney (1995) and Leigh et al. (2004) observed that eustatic effects are not apparent more than 60 to 80 km inland from the modern shoreline in the Ogeechee River valley and the Altamaha and Pee Dee River valleys, respectively, and are limited to thin veneers of Holocene sediments deposited during sea level rise and valley backfilling that onlap and bury Pleistocene terraces graded to the slightly lower level of the formerly exposed continental shelf. Similar upstream limits of eustatic effects during the late Pleistocene have been reported by Blum and Aslan (2006) and Otvos (2005) in the Gulf Coastal Plain and are thought to be related to climate driven terrestrial sediment yield that overwhelmed and precluded upstream incision induced by lower sea level and restricted it to the outermost continental shelf and continental slope (Blum and Aslan, 2006; Leigh, 2008). The Little River, located some 130 km inland from the head of the estuary of the modern Cape Fear River, would have been insulated from base level effects during the late Pleistocene. This reasoning is consistent with the finding that the lowest incision rate calculated for the Little River (0.07 mm/yr) during the late Pleistocene spans the period of base level lowering associated with the last glacial maximum, between the deposition of T3b and T2 (59.7 to 17.4 ± 4.2 ka).

Recent studies have documented climate-driven changes in channel patterns and sedimentation styles in rivers of the southeastern Atlantic Coastal Plain (Leigh, 2008; Leigh, 2006; Leigh et al., 2004), and Little River data may indicate that environmental change influenced incision during the late Pleistocene. Net incision rates based on mean ages between the deposition of T3b and T2 (0.07 mm/yr, 59.7 to 17.4 ka, late OIS 4 to OIS 2) are comparatively lower than the rates between the deposition of T5b and T4 (0.65 mm/yr, 94.0 to 74.6 ka, OIS 5b to early OIS 4), T4 and T3b (0.50 mm/yr, early through late OIS 4), and T2 and T1 (0.56 mm/yr, 17.4 to 10.2 ka, OIS 2 to 1) (Table 4). The timing of possible variation in Little River incision shows a general correspondence to late Pleistocene climate-driven changes in river regime documented elsewhere in the southeastern Coastal Plain (Leigh, 2008). Leigh et al. (2004) documented braided channel patterns on terraces along seven Coastal Plain rivers, including the Cape Fear River. Optimal expression of braiding occurred between 17 – 30 ka (OIS 2), with braiding possibly occurring as early as 69 ka (OIS 4). Paleomeanders on terraces immediately predating braided terraces in the Pee Dee and Altamaha River valleys indicate rivers had meandering planforms prior to OIS 4 (Leigh et al., 2004; Leigh, 2008). Leigh et al. argue that braiding resulted from high sediment loads and bank instability related to the cooler, drier climatic conditions during OIS 2 through 4 that may have had a snowmelt runoff season that caused larger bankfull discharges than present. Channel patterns transitioned from braided to meandering at about 15 – 16 ka in response to global warming and regional increases in moisture that promoted a more densely vegetated landscape that resulted in reduced sediment yield and increased resistance of channel bank materials (Leigh, 2006).

Although Leigh et al. (2004) and Leigh (2006) do not present incision rates for Coastal Plain rivers during the late Pleistocene, given that Coastal Plain rivers appear to have been sandy, aggrading, braided systems during OIS 4 through 2, it is reasonable to suspect they would exhibit relatively low incision rates during this period. Little River incision rates during OIS 4 through 2 are comparatively lower than those exhibited during OIS 5 to 4 and OIS 2 to 1 and are consistent with this interpretation. Rapid incision to the elevation of the modern channel bed that occurred between the deposition of T2 to T1 (0.56 mm/yr, 17.4 to 10.2 ka) roughly coincides with the braided to meandering transition documented at about 15 – 16 ka for other Coastal Plain rivers that was followed by downcutting to the level of modern floodplains (Leigh, 2006).

Interestingly, incision accompanying terminal Pleistocene transitions from braided to meandering channels has also been documented for unglaciated catchments in the upper mid-latitudes of central and western Europe. Kozarski (1991) attributed the braided to meandering transition and associated downcutting of the Warta River (Poland) ca. 13,000 ^{14}C yr BP to decreases in sediment yield, bank erodibility, and stream power that resulted from climate-driven expansion of vegetation cover and changes in the river's discharge regime. Starkel (1991) also implicated climate-mediated changes in hydrologic regime and sediment supply as contributing factors to similar planform changes and downcutting that occurred along the upper Vistula River (Poland) ca. 13,000 ^{14}C yr BP but noted that incision was in part driven by lowering of base level following retreat of the Scandanavian ice sheet from the lower course of the Vistula's modern drainage. Similar climate-induced changes to sediment: discharge relationships also explain incision beneath the last glacial braidplain of the Maas River of northwestern Europe (Vandenberghe et al., 1994). In the above cases, several intervals of incision and aggradation, accompanied by transitional planform stages, have been documented between late Pleistocene braided systems and the establishment of Holocene systems with modern bed elevations and high sinuosity, modern-size meanders. Such transitional phases are not apparent in the Little River record, at least at its present resolution.

Little River incision between deposition of T5b and T4 (0.65 mm/yr, 94.0 to 74.6 ka, OIS 5b to OIS 4) may indicate river response to the climate of the last interglacial, when rivers may have had meandering patterns (Leigh et al., 2004) and when the climate of the southeastern United States was warmer and wetter than during the Wisconsinan (Watts, 1971). The relatively high incision rate between OIS 5 and 4 may reflect river response to more sedimentpoor waters that were the result of lower sediment yields from a landscape that was more densely vegetated during the last interglacial than during OIS 4 through 2. Although unequivocal examples of incision in response to last interglacial climatic conditions have not been documented for other rivers of the southeastern Atlantic Coastal Plain, incision during this time has been reported for some European-rivers. Straffin et al. (1999) observed 15 m of vertical incision within a dated terrace sequence along the Loire River (western France) that they attributed to a reduction in sediment supply due to more moist, densely vegetated landscape conditions during the last interglacial. Along the Somme River of northern France, Antoine (1994) reports incision occurring somewhat later, between OIS 5e and 4, during transitional interglacial to early-glacial time, while the fluvial record of the Mediterranean region is more complex, with apparent climate-driven incision occurring during OIS 5e followed by periods of aggradation during parts of OIS 5d (~109-111 ka) and around the OIS 5b/5a boundary (~88 ka) (Macklin et al., 2002).

Human impacts or intrinsic mechanisms within the fluvial system (Schumm, 1973) also may cause rivers to incise. Because widespread human occupation of the southeastern Atlantic Coastal Plain did not occur until the terminal Pleistocene (Hoeffecker et al., 1993; Meltzer, 1988) and Native American land use practices were probably insufficient to force major changes within Coastal Plain rivers (Leigh, 2008), human impacts can be ruled out as a driver for Little River incision during the late Pleistocene. Effects of intrinsic mechanisms, thresholds, and complex response are difficult to assess with the available data, and some internal controls on incision may have operated on the Little River. However, rough correspondence between incision rates based on the statistical means of age estimates and documented phases of river response to paleoclimatic variation in the region may indicate Little River terraces reflect climate-driven aggradation compensated by incision in response to slow, gradual neotectonic uplift.

CONCLUSIONS

The Little River valley contains a floodplain and at least five mappable terraces that were deposited within the last 94.0 ±15.9 ka. During this time, the net incision of the Little River was 27 m, with a long-term rate of 0.29 mm/yr. This value is consistent with incision rates calculated by the authors from terrace age and elevation data from late Pleistocene and Holocene terraces along other rivers in the eastern United States compiled by Mills (2000), but it is nearly an order of magnitude higher than the incision rate for the northwestern portion of the Cape Fear River valley reported by Soller (1988) during the late Pleistocene based on a 90 ka age estimate for the Wando terrace. Data are insufficient to allow correlation of Little River and Cape Fear River terraces, and differences in the fluvial systematics and external and internal drivers of incision and aggradation among the Little River and other southeastern rivers during the past 100 ka limit the ability to extrapolate the findings of this study to other rivers in the region. However, age estimates from the Wando terrace in the Pee Dee River valley published elsewhere (Leigh et al., 2004; Thom, 1967) are consistent with the long-term Little River incision rate and together these data suggest the fluvial facies of the 90 ka Wando Formation, which is mapped by Owens (1989) as the lowest terrace above the Holocene floodplain in the Cape Fear and Pee Dee River valleys, is actually comprised of multiple fluvial deposits that span a wide range of late Pleistocene ages.

The long-term net incision rate of the Little River is within the range of neotectonic uplift rates reported for the southeastern Atlantic Coastal Plain, and the variation in incision rates calculated from inter-terrace mean ages shows a general correspondence to the timing of late Pleistocene climate-driven changes in river regime reported elsewhere in the region. Although insufficient temporal resolution of terrace ages precludes definitive conclusions about causes and variation of incision during the late Pleistocene, Little River terraces may reflect short-term aggradation in response to periodic climate-mediated increases in sediment supply that occurred within the overall context of long-term incision in response to neotectonic uplift.

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A New Species of Schizaster (Echinoidea, Spatangoida) from the Late Pliocene (Placenzian) Intracoastal Formation of Liberty County, Florida

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ABSTRACT

A new species of spatangoid echinoid, *Schizaster kieri* n. sp., from the late Pliocene Intracoastal Formation, Liberty County, Florida is described and discussed. Schizaster *kieri* n. sp., is the first species of the genus *Schizaster* described from the Pliocene of the east coast of North America. It is readily differentiated from it's stratigraphically nearest eastern North American congener, the Oligocene age *Schizaster americanus* (Clark), by its much more elongate test and more posterior apical system, among other traits. The addition of *Schizaster kieri* n. sp., to the echinoid fauna of the Intracoastal Formation, increases the known echinoid fauna of the formation to nine species.

INTRODUCTION

The echinoid fauna of the Intracoastal Formation in the south-central Florida panhandle has been little studied primarily due to the subsurface position of the unit. However, over the past decade, quarrying operations north of Carrabelle, in southeastern Liberty County, Florida, have offered an unprecedented opportunity to examine and collect the Intracoastal Formation. Oyen (2001), in his unpublished master's thesis, lists *Clypeaster* sp., *Encope aberrans* (Martens, 1867) and *Echinocardium orthonotum* (Conrad, 1843) as occurring in the echinoid fauna of the Intracoastal Formation. This was later formalized by Oyen and Portell (2001), which constituted the first formal reference to the echinoid fauna of the Intracoastal Formation. Collecting by the authors has provided a more complete picture of the echinoid fauna of this unit (Osborn and Ciampaglio, 2010).

Specimens of *Schizaster kieri* n. sp., were collected from a highly fossiliferous, Upper Pliocene, three meter thick, tan-gray, poorly consolidated, sandy biocalcarenite facies within the Intracoastal Formation located within an active quarry in southeastern Liberty County, Florida. In this horizon, the species rarely occurs, but is part of a diverse assemblage of echinoids, including *Arbacia improcera* (Conrad, 1843), *Eucidaris* aff. *tribuloides* (Lamarck, 1816), *Clypeaster sunnilandensis* (Kier, 1963), *Echinocardium orthonotum* (Conrad, 1843), *Plagiobrissus sarae* (Ciampaglio, et al., 2009), and a small undescribed species of *Genocidaris* currently being studied by the authors (Osborn and Ciampaglio, 2010). Specimens of an undescribed species of the Raninidae crab genus *Ranilia* are notably abundant within, and are characteristic of, this horizon (Portell, et. al., 2003).

Overlying the *Ranilia* biozone, within the study area, is a one meter thick, shell rich, sandy calcarenitic limestone of the upper Intracoastal Formation which contains specimens of the echinoid *Encope macrophora* (Ravenel, 1842) and *Encope aberrans* (Martens, 1867) but lacks *Echinocardium orthonotum* (Conrad, 1843), *Arbacia improcera* (Conrad, 1843) and *Schizaster kieri* n. sp., which are the characteristic echinoidea of the underlying *Ranilia*-biozone biocalcarenites.

The echinoid fauna of the Intracoastal Formation is closely allied to that of the partially correlative upper Pliocene Tamiami Formation (Figure 1) of the southern Florida peninsula. Both units contain the species *Eucidaris* aff. *tribuloides* (Lamarck, 1816), *Arbacia improcera* (Conrad, 1843), *Clypeaster sunnilanden-*

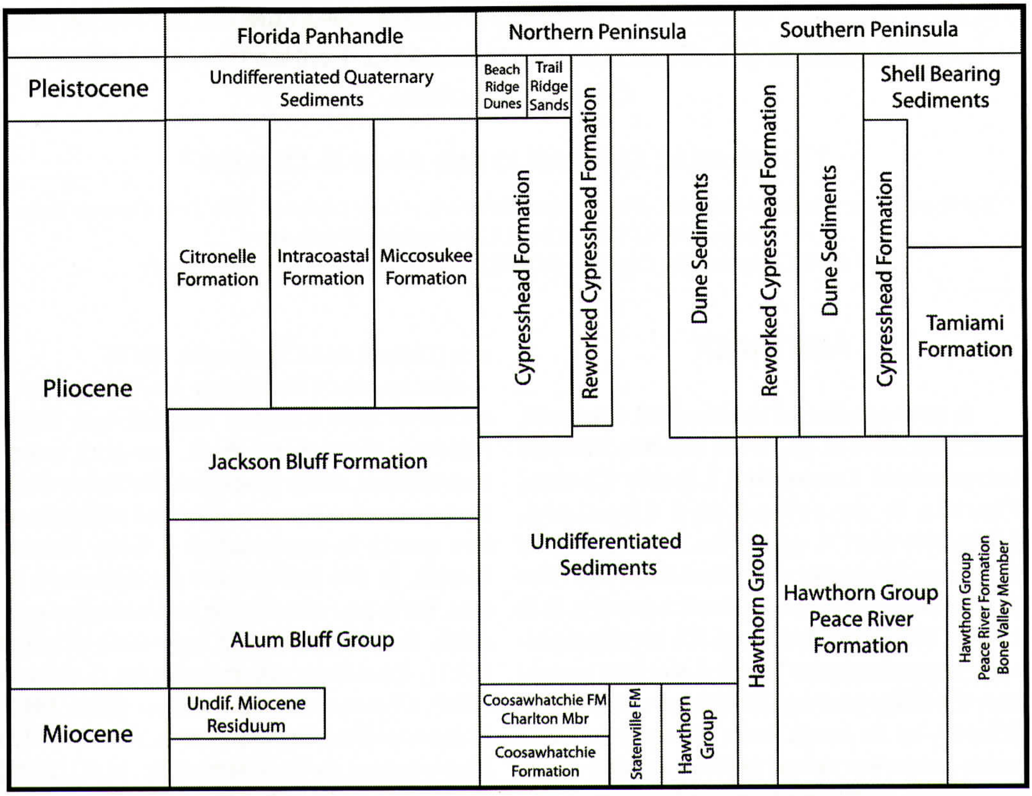

Figure 1: Stratigraphic column for the Pliocene of the Florida, (from Scott, 2001).

sis (Kier, 1963), *Echinocardium orthonotum* (Conrad, 1843) and *Plagiobrissus sarae* (Ciampaglio, et al., 2009) (Figure 2). Though the genus *Plagiobrissus* has not been formerly documented in the Tamiami Formation, recent collecting within this formation in the southern Florida peninsula has provided specimens of *Plagiobrissus sarae* (Ciampaglio, et al., 2009) for study. The occurrence of *Plagiobrissus sarae* (Ciampaglio, et al., 2009) in the Tamiami Formation, coupled with the presence of *P. sarae* in the Intracoastal Formation discussed within this paper, greatly expands the distribution of this species southward from the coastal plain of South Carolina (Ciampaglio, et al., 2009). The genus *Schizaster* has not been recorded from the Tamiami Formation (Clark and Twitchell, 1915; Cooke, 1959; Kier, 1963; Mansfield, 1932; Oyen, 2001; Oyen and Portell, 2001).

GEOLOGIC SETTING

Schizaster kieri n. sp., specimens described herein were collected from the late Pliocene (Piacenzian), Intracoastal Formation, in Liberty County, Florida. The specimens were obtained from a *Ranilia* biozone that is roughly three meters in thickness, and resides one meter below the top of the Intracoastal Formation section in the study area.

The Intracoastal Formation was first described by Huddleston (1976), and takes its name from the Intracoastal Waterway #1 core, located in Walton County, Florida (Schmidt, 1984). The unit is restricted to the Apalachicola Embayment of the south central Florida panhandle, and is upper Pliocene in age (Scott 2001) (Figure 1). Schmidt (1984) discussed a middle Miocene age for the lower portion of the formation that is separated from the late Pliocene age, upper portion of the formation, by an unconformity. Schmidt's (1984) division of the

Figure 2: Echinoid fauna correlation of significant upper Pliocene age echinoid-bearing formations of southeastern North America, from Cooke (1959), Kier (1963), Oyen (2001) and extensive collecting by the authors.

Species	Intracoastal Formation	Tamiami Formation	Goose Creek Limestone	Notes
Eucidaris tribuloides	X	X		New record for the Intracoastal Formation
Genocidaris sp.	X			Undescribed species
Lytechinus cf. variegatus			X	
Lytechinus variegates pluri-tuberculatus		X		
Arbacia improcera	X	X	X	Synonymy *Arbacia crenulata* (Kier 1963)
Arbacia rivuli			X	
Arbacia sp.			X	Undescribed species
Arbacia waccamaw			X	
Clypeaster romani		X		Synonymy *C. crassus* (Kier 1963); see Kier 1964
Clypeaster sunnilandensis	X	X		New record for the Intracoastal Formation
Mellita aclinensis		X		
Mellita caroliniana			X	
Encope aberrans	X			
Encope aberrans imperforata		X		Syn *E. michelini imperforata* (Kier 1963); see Phelan 1972
Encope macrophora	X		X	
Encope tamiamiensis		X		
Rhyncholampas evergladensis		X		
Rhyncholampas sabistonensis			X	
Schizaster kieri n. sp.,	X			
Agassizia porifera		X	X	
Echinocardium orthonotum	X	X	X	Synonymy: *Echinocardium gothicum* (Ravenel 1848)
Spatangus glenni			X	
Brissus glenni			X	
Brissus cf. glenni		X		New record for the Tamiami Formation
Brissus cf. unicolor			X	
Plagiobrissus sarae	X	X	X	New record for the Tamiami and Intracoastal FMs

Intracoastal Formation has not been disputed as of the writing of this paper, though it was not reflected in the most recent geological map of Florida (Scott, 2001). The echinoid fauna discussed herein was collected from the upper Pliocene age biocalcarenites near the top of the formation.

The Intracoastal Formation is an extremely fossiliferous formation which contains a highly diverse micro and mega fauna. The formation consists of a very sandy, poorly consolidated argillaceous limestone that is generally referred to as a poorly consolidated wackestone or biomicrite (Schmidt, 1984). The formation thickens and dips to the south-southwest, and approaches 100 feet in thickness in the southeastern corner of Liberty County, near the collecting area (Rupert 1991). The diverse group of marine organisms, as well as the significant amount of quartz sand, heavy minerals, clays, glauconite and phosphate lends to the conclusion that the Intracoastal Formation was deposited on a shallow-shelf, supplied by fluvial sources (Schmidt, 1984).

Within the study area of southeastern Liberty County, the Intracoastal Formation is overlain by undifferentiated Quaternary quartz sands and clay (Rupert, 1993). Elsewhere within the Apalachicola Embayment, the formation is more frequently overlain by the upper Pliocene Jackson Bluff Formation (Rupert, 1991; Schmidt, 1984); however, Scott (2001) reports the Intracoastal Formation as overlying the Jackson Bluff Formation (Figure 1). This placement of the formation stratigraphically above the Jackson Bluff Formation conflicts with previously published data (Rupert, 1991; Rupert 1993; and Schmidt, 1984), and should be revisited.

SYSTEMATIC PALEONTOLOGY:

Figured specimens of *Schizaster kieri* n. sp. are housed at the North Carolina Museum of Natural Sciences (NCSM) in Raleigh.

Class ECHINOIDEA Leske, 1778

Order SPATANGOIDA Claus, 1876

Suborder PALEOPNEUSTINA Markov and Solovjev, 2001

Family SCHIZASTERIDAE Lambert, 1905

Type genus *Schizaster* Agassiz, 1836

Type species *Schizaster studeri* Agassiz, 1836

Schizaster kieri Osborn, new species

(Figure 1, Plate 1)

Diagnosis: Test elongate, sharply wedge-shaped profile, deeply sunken ambulacra, ambularca I and V 44% the length of ambulacra II and IV, with a pronounced, deep anterior sulcus. Apical system has four genital pores: anterior pair are significantly smaller than the posterior pair.

Description: Test is ovate, wedge shaped, widest point slightly forward of apical system, with a prominent deep anterior sulcus. A sharp keel extends on aboral surface from the apical system to the posterior margin. Aboral surface slopes at approximately twenty degrees from the apex, posterior of the apical system, toward the anterior margin. Aboral tuberculation is fine and dense. Apical is posterior of center, 60% of test length from the anterior margin of the test, and 40% of test length from the posterior margin with four gonopores, anterior pair half the size of posterior pair. Ambulacrum III is deeply sunken aborally, 59% of test length on all measured specimens (Table 1), and forms sunken furrow, and distinct notch in anterior margin of test; not sunken at peristome, with 44 small, conjugate, uniserial pore pairs in holotype. Length of ambulacra II and IV is 40% of test length on holotype; with 34 pore pairs; moderately sunken with posterior curve; divergent at a 40 degree angle. Ambulacra I and V are straight, shorter, 18% of test length in holotype, 44% the length of ambulacra II and IV in all examined specimens (Table 1); 17 pore pairs in holotype. Ambulacra and interambulacra extend to peristome. Peristome situated anteriorly,

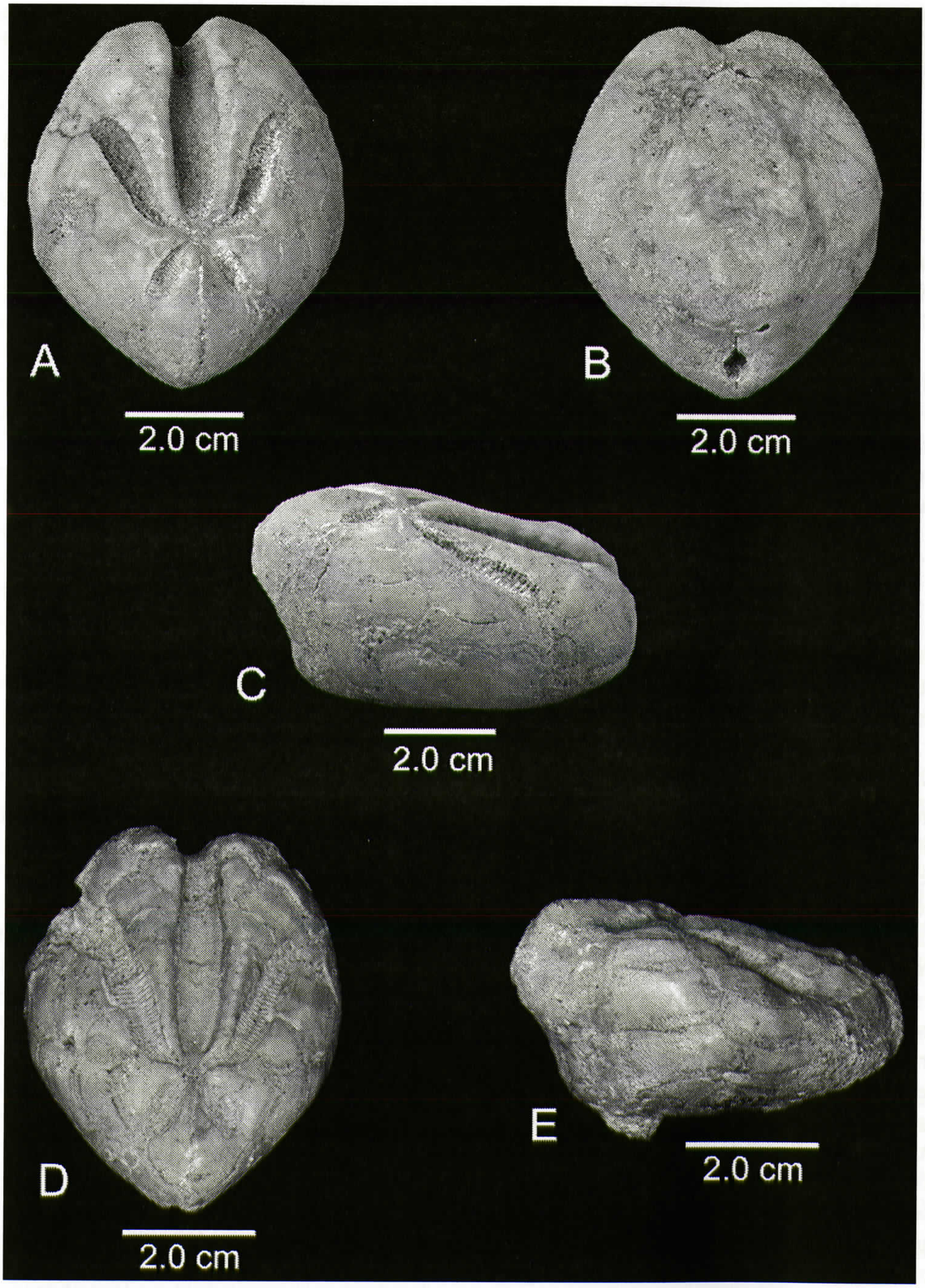

Plate one: Type Specimens of *Schizaster kieri* n. sp. Type specimens are housed at the North Carolina Museum of Natural Science.
A: *Schizaster kieri*, NCSM 11397, holotype, aboral view, Liberty County, FL, 72mm x 59mm.
B: *Schizaster kieri*, NCSM 11397, holotype, oral view, Liberty County, FL, 72mm x 59mm.
C: *Schizaster kieri*, NCSM 11397, holotype, lateral view, Liberty County, FL, 72mm x 59mm.
D: *Schizaster kieri*, NCSM 11398, paratype, aboral view, Liberty County, FL, 58mm x 48mm.
E: *Schizaster kieri*, NCSM 11398, paratype, lateral view, Liberty County, FL, 58mm x 48mm.

Table 1: Measurements of *Schizaster kieri* n. sp. Indeterminable measurements are denoted by NA due to test preservation or incompleteness. All measurements are in millimeters. Non-type specimens reside in the private collections. (amb=ambulacrum)

Character	Holotype NCSM11397	Paratype NCSM11398	Non paratype
Length	71.6	58.1	53.1
Width	58.9	48.1	44.7
Height at anterior margin	21.0	NA	12.7
Height at Posterior margin	40.6	NA	34.4
L amb I	12.6	10.6	9.2
L Amb II	28.4	23.9	20.8
L Amb III	42.1	34.4	31.7
L Amb IV	26.7	NA	20.8
L Amb V	12.6	10.6	9.2
Periproct Width	5.2	7.3	4.7
Periproct Height	6.8	NA	5.4
Peristome Width	8.7	NA	4.8
Porepairs Amb I	17	18	15
Porepairs Amb II	34	36	30
Porepairs Amb III	44	NA	38
Porepairs Amb IV	34	NA	30
Porepairs Amb V	17	NA	15

the distance from posterior end of test to lip of labrum of holotype is 48mm. Labrum well developed, short, width of 8mm in holotype; plastron 30mm wide at point of greatest width. Latero-anal fasciole makes nearly straight line from peripetalous fasciole to posterior end of test. Peripetalous fasciole broadest at end of ambulacra, uniform in width, crosses ambulacrum III 33 mm from apical in holotype. Periproct ovoid, vertical diameter 6.8mm; horizontal diameter 5.2mm, situated on vertical truncated posterior.

Discussion: *Schizaster kieri* n. sp., is the first member of the genus *Schizaster* described from the Pliocene of the east coast of North America. It is readily differentiated from it's stratigraphically nearest eastern North American congener, the Oligocene age *Schizaster americanus* (Clark), by its much more elongate test, greater wedge-shaped profile and more posterior apical system, among other traits.

Smith (et al., 2005) synonymized *Schizaster* and *Paraster* based in part on the fact that the apical system of the type specimen of *Schizaster studeri* (Aggasiz 1836) was not sufficiently preserved to display the genital pores. Mortensen (1951) based his classification of the genus on a statement by Lambert that the holotype had two genital pores, even though Cotteau suggest that it had four. Mortensen (1951) then mistakenly separated *Schizaster* from *Paraster* based on the number of genital pores, with *Schizaster* having two and *Paraster* having four. Smith (et al., 2005) stated that a well preserved specimen of *Schizaster studeri* (Aggasiz 1836) from the type locality has four genital pores. Therefore, they determined that Mortensen's consideration of *Paraster* and *Schizaster* as separate genera, based on the number of genital pores is erroneous, and the two are clearly synonymous. The authors agree with this determination, therefore the specimen described herein is attributed to the genus *Schizaster* (Agassiz, 1836) which has precedence over *Paraster* (Pomel, 1869).

Schizaster kieri n. sp. is readily differentiated from the three recent members of Schizasteridae that inhabit Caribbean and Gulf of Mexico waters: *Schizaster orbignyanus* (Aggasiz 1880), *Schizaster floridiensis* (Kier and Grant, 1965; Chesher, 1966), and *Schizaster doederleini* (Chesher 1972), by its much more elongate test, deeper anterior sulcus, and greater wedge-shaped profile. *S. kieri* n. sp., is differentiated from *S. orbignyanus* (Aggasiz 1880) by the fact that *S. orbignyanus* has merely two genital pores, and should be attributed to the genus *Ova*.

Pliocene age *Schizaster* are rare in the Caribbean region; however, three species have been documented; Kier (1984) and Sanchez Roig (1949) discuss *Schizaster cubensis* (D'Orbigny 1842), of the Vedado Habana province of Cuba, and attribute it to the Pliocene. Though *S. cubensis* is present merely as partial specimens, *Schizaster kieri* n. sp., can be differentiated from it by its more elongate test, sharply wedge shaped profile and more depressed ambulacra. Additionally, Engel (1961) described *Schizaster eustatii*, from the Pliocene of St. Eustatius. *Schizaster kieri* n. sp., is differentiated from *Schizaster eustatii* (Engel 1961) by its deeper ambulacra, more elongate test, and much more pronounced wedge-shaped profile. Kier (1992) reports the occurrence of the extant species *Schizaster doederleini* (Chesher 1972) in the Pliocene of the Dominican Republic. However, *Schizaster kieri* n. sp., is differentiated from the much more rotund *S. doederleini*, by its much more elongated test, deeper anterior sulcus, and greater wedge-shaped profile.

Donovan and Portell (1998) discuss incomplete and compressed specimens of the genus *Schizaster* from the upper Pliocene age Bowden Formation of southeast Jamaica. However, the preservation of the specimens was insufficient for them to make conclusions concerning the specific identity of the specimens. Additional study and more complete specimens would be required to determine if these specimens should be referred to *Schizaster kieri* n. sp.

Material: Holotype NCSM11397 and paratype NCSM11398 (Plate 1). Material examined includes the holotype, paratype, and one other less detailed and complete non-type specimen, collected by the authors in Liberty County, Florida.

Measurements: Measurements for all examined specimens are given in table 1

Etymology. This species is named in honor of Porter Kier, in recognition of his extensive work detailing the Cenozoic echinoid faunas of the eastern United States and Caribbean, and his tremendous contributions to echinoid paleobiology.

Occurrence. *Schizaster kieri* n. sp., has not been documented outside of Liberty County, Florida where it rarely occurs in the late Pliocene age Intracoastal Formation in a quarry north of Carrabelle, east of Route 67, just north of the Liberty County line.

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